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Stratigraphic Investigations Based on Eustatic Fluctuations

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Abstract – One trend in stratigraphy, Sequence Stratigraphy (SS), has a fundamental objective: stratification systems. Their formation is related to a low sea-level, followed by a transgression, and finally a high sea level. The application of stratification systems to stratigraphic interpretation of sedimentary basins is discussed. Restrictions in the employment of SS are presented, and the sites where SS may be applied are defined.

Sequence Stratigraphy (SS), a new trend in studying sedimentary basins, was developed over the past 15 years. Global sea-level changes and procedures of the seismostratigraphic analysis are the basis of SS. Event Stratigraphy, which has been developing rapidly since the mid-1980s, incorporates SS as an important component (Il'in, 1991). SS has attracted the attention of many scientists, a fact confirmed at the International Congress on Sedimentology, held in Nottingham in 1990, where SS was one of the leading sections. SS, which uses the latest seismic survey and relies upon the newest data on sedimentation processes in the seas and oceans, is developing quickly because it aids the process of locating and mining natural oil and gas reservoirs. Many deposits of hydrocarbons (mainly on shelves) have already been discovered with the help of SS.

The SS technique has been widely used in investigations abroad since the 1970s. In well-known publications, P.R. Vail, N. Christie-Blick, B.U. Haq, and many other researchers, using data on various regions of the world for their conclusions, clarified the principles of identifying both the depositional sequences and their inner structures (Christie-Blick *et al.*, 1988; Haq *et al.*, 1987, 1988; Mitchum and Van Wagoner, 1991; Seismic Stratigraphy..., 1977; Vail, 1987; Vail *et al.*, 1991; Van Wagoner *et al.*, 1987).

In Russia, SS was elaborated upon to a considerably lesser degree due to, among other reasons, the weaker resolution of our seismic prospecting techniques, a lack of scientific teams incorporating highly skilled specialists from different institutions, and a limited number of sites suitable for such kinds of investigations.

In Russian scientific literature, the main postulates concerning SS were presented by K.R. Volkov and A.E. Shlezinger (1992). Information on the International Congress in Nottingham, dedicated to SS, was presented by A.V. Il'in (1991). During the past 2 to 3 years, SS accumulated much new data. This paper deals with the present state of SS and with problems raised in recent publications.

Traditional stratigraphy assumes that paleontological data is the basis of the subdivision of Phanerozoic geological sequences by identifying chronostratigraphic units like stages and zones. SS analyzes the sedimentary sequences genetically related to global sea-level fluctuations. The sequences are bounded by unconformities defined by seismic characteristics in well logs and in natural outcrops. With sufficiently complete sections, the unconformities can be dated biostratigraphically. SS gives special consideration to the study of rock relationships within sedimentary bodies that are also bounded by the surfaces of breaks in the sedimentation and by erosion, of a higher order, or else by time-corresponding boundaries of conformable strata.

With this approach, the fundamental stratigraphic unit is a sedimentary sequence that represents a three-dimensional geological body bounded by unconformities or, sometimes, by corresponding conformable surfaces. The sequence corresponds to a major sedimentary stage (from 1 or 2 up to 10 Ma), and is composed of genetically related strata. It can be subdivided into sedimentary systems with different facies compositions. Such systems can be subdivided into subsystems and are bounded by surfaces produced by sea-level fluctuations. The boundaries of the systems and subsystems provide a chronostratigraphic framework for correlating and mapping the sedimentary rocks. The systems and subsystems are defined and identified by their lateral distribution, their bounding surfaces geometry, and the shapes of their sedimentary bodies. The thickness of the units, the time intervals during which they were formed, and the interpretation of their origin are not used to identify them (Figs. 1 - 4). The sedimentary (seismic) sequences were defined both in our country and abroad before SS appeared; but those sequences did not bear the genetic meaning which SS gave to them, i.e., their formation was not seen as being related to eustatic fluctuations.

At the present time, it is increasingly evident that sedimentary sequences, and the sedimentary systems

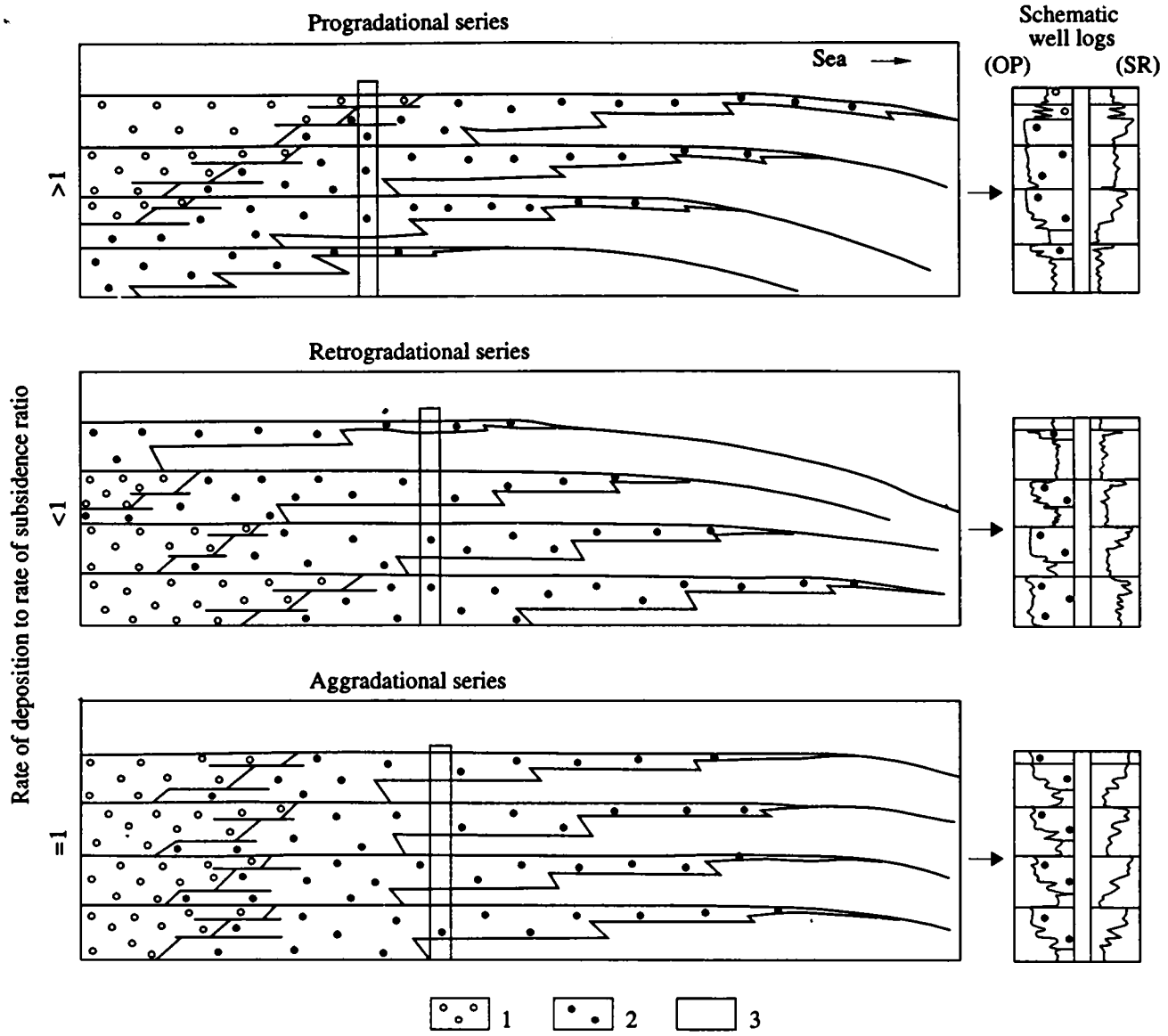

Fig. 1. Facies pattern in stratification systems (Van Wagoner *et al.*, 1987). 1 – Coastal-plain sandstones and claystones; 2 – shallow-water marine sandstones; 3 – shelf claystones.

which compose them, reflect three main geological processes: (1) sea-level fluctuation, (2) subsidence and the uplifting of the basin floor caused by tectonic activity, and (3) a considerable sediment supply entering into the basin. The sedimentary sequences are the main units used to elaborate the stratigraphic scale. Within this scale the chronostratigraphic or lithofacies correlation can be done. These sequences can be identified by seismic and drilling data and by their natural outcrops. Facies succession within sequences may be progradational (in the case of a seaward-moving shoreline), retrogradational (for a landward-moving shoreline), and agradational (for a stable shoreline) depending on the ratio of the sedimentation rate to the rate of sea-floor subsidence (Fig. 1). Unconformities bounding the sedimentary sequences separate the older and younger strata. The former are often considerably eroded. Unconformities form at a low sea level and correspond to a point of a bend in the eustatic curve. There are unconformities related to subaerial erosion (the first type) and those related to subaqueous erosion (the second type). The unconformities of the first type are formed when the rate of the sea-level drop exceeds that of the basin-floor subsidence. On the contrary, unconformities of the second type are formed when the rate of the drop is less than the rate of subsidence, and the sea floor remains under water. In this case, erosion seems to be caused by a subsequent transgression, which creates a new level of the shelf. This process was studied in detail by N.V. Esin *et al.* (1986). Local erosion, concurrent with the deposition of shoal and migrating dune sediments, is not included in the unconformities. In a deep-water environment, underwater erosion is associated with submarine

currents whose origin depends on climatic changes and sea-level fluctuations.

As was stated above, the sedimentary sequences are identified on the basis of seismic and well-log data and natural outcrops. Sometimes their boundaries are difficult to define, as they are usually confined to the inner parts of regressive units in the section. The base of the sequence can be recognized by a sharp seaward displacement of sedimentation and by obvious unconformities. In places without unconformity, the boundary may be identified by the replacement of progradational beds by aggradational ones. Sedimentary sequence boundaries, as well as unconformities within the sequences, are recognized in a seismic record by identifying terminations of seismic reflections in base discordance, baselap, toplap, and top truncation. If strata are parallel, surfaces of unconformity are defined according to biostratigraphic data.

Two types of sedimentary sequences may be recognized, depending on the characteristics of their bases (Figs. 2 - 4). The first is bounded from below by a subaerial erosion surface. The second is bounded by a subaqueous erosion, or conformable, surface. The type-1 sequences are composed of low sea-level, transgression, and high sea-level stratification systems. The type-2 sequences comprise outer-shelf (shelf-margin), transgression, and high sea-level stratification systems. The high and low sea-levels are of relative value and are not confined to specific cycles of eustatic or relative sea-level fluctuations.

The base of a type-1 sequence is a buried paleolandform with a relief ranging from tens of centimeters to a few metres. The paleotopography level has age analogs on the coastal plain and in the basin. In these areas, surfaces synchronous to the paleolandform can be identified by correlating the underlying and overlying strata.

The low sea-level stratification system of the type-1 sequence rests on the surface of the underlying sequence and pinches out at the paleoshoreline. It is made up of a basin-floor fan, a slope fan, clinoforms (prograding wedges), and a fill of incised valleys. The basin-floor fan consists of low, elongated mounds composed of cherty and clastic sediments with prevailing coarse sands and carbonate debris. This material passes across a shelf and a continental slope, being transported via submarine valleys and canyons to the basin floor. The slope fan is made up of turbidites, or the deposits of gravity flows. It may form simultaneously with a basin-floor fan or with an initial stage of a clinoform

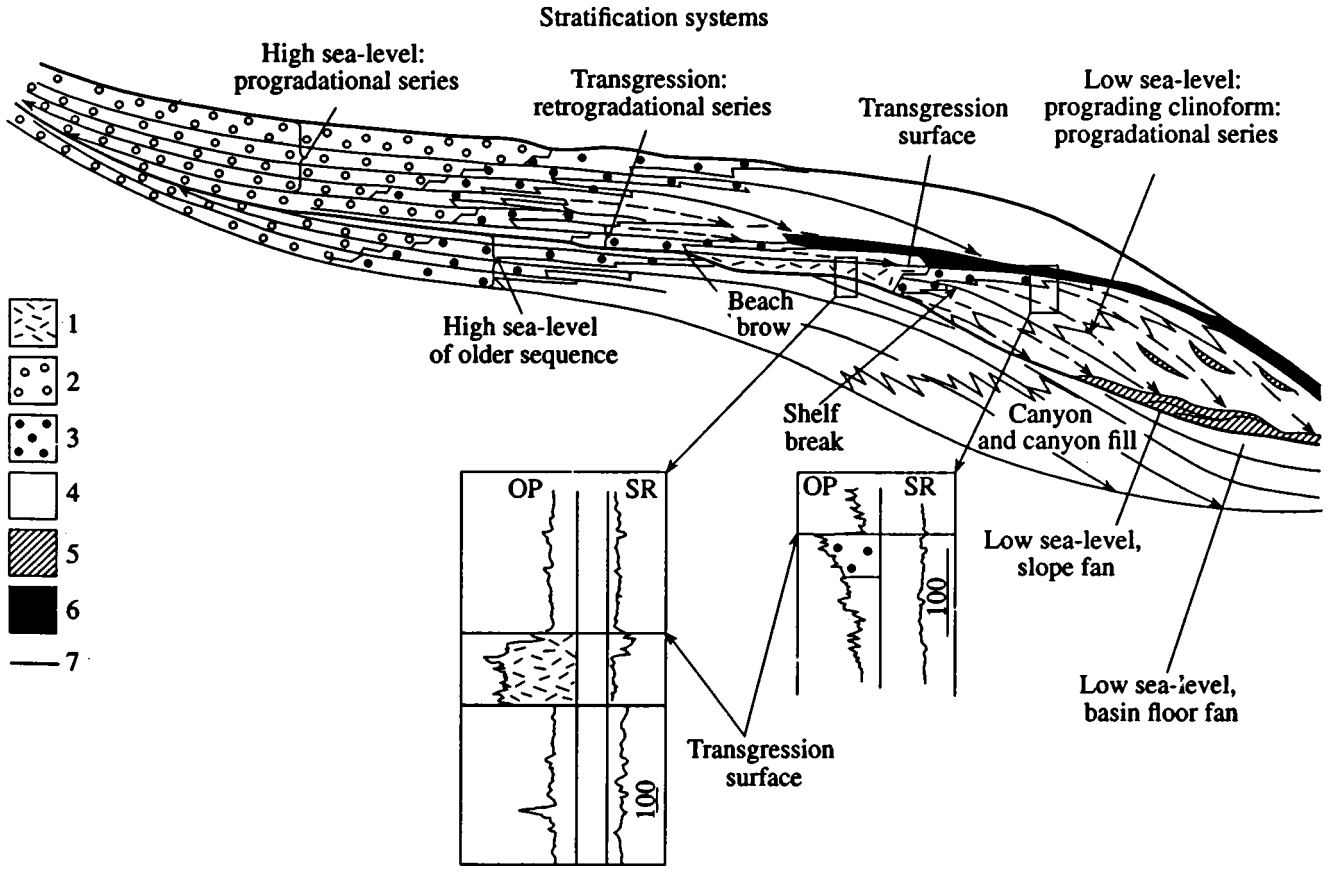

Fig. 2. Stratification systems in a type-1 sequence within the shelf, slope, and basin (Van Wagoner *et al.*, 1987). 1 – Channel and estuarine sandstones within incised valleys; 2 – coastal-plain sandstones and claystones; 3 – shallow-water marine sandstones; 4 – shelf and slope claystones and thin sandstone beds; 5 – submarine fan sandstones; 6 – condensed deposits; 7 – type-1 sequence base.

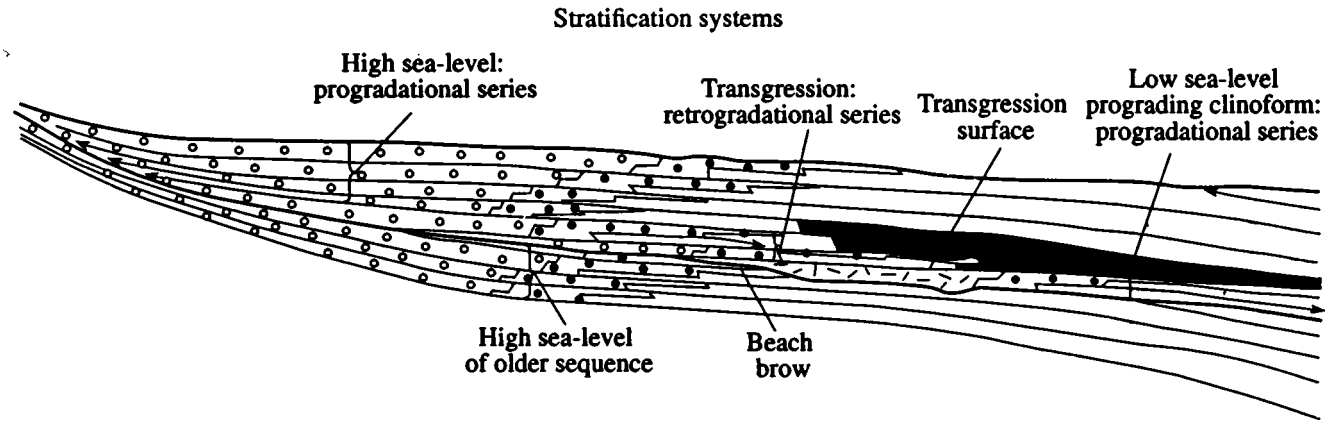

Fig. 3. Stratification systems of a type-1 sequence on a shelf (Van Wagoner *et al.*, 1987) (for explanation see Fig. 2).

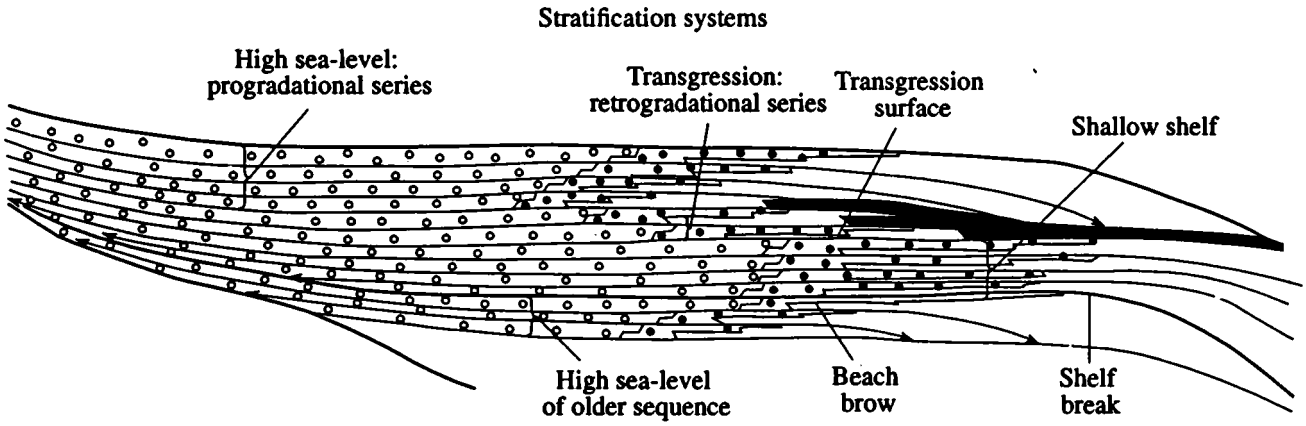

Fig. 4. Stratification systems in a type-2 sequence (Van Wagoner *et al.*, 1987) (for explanation see Fig. 2).

growth. The clinoforms are composed of strata, which vary in grade from progradational to aggradational. The terrigenous material they are composed of becomes finer upsection. They bury the incised valleys in a paleoshelf. Below the shelf break, the middle and upper parts of the clinoform overlie the roof of the slope and the basin-floor fans with a baselap. When the slope and basin are absent, the low sea-level stratification systems consist of lower and upper clinoforms. The latter bury the incised valleys.

The outer paleoshelf stratification system of the type-2 sequence is composed of weakly progradational to aggradational clinoform strata. Characteristic of their base are a landward base discordance and a seaward baselap. The paleoshelf stratification system successively progrades toward the outer sea and buries the slope.

The transgression stratification system occupies the central sections of both type-1 and type-2 sequences. It is characterized by progradational strata. Landward, the strata of the transgression stratification system overlie the sequence boundary, and seaward they lie upon the underlying strata of the low sea-level system through a baselap and become thinner until they are completely pinched out in some places. The transgression stratification system is often composed of condensed (depression) deposits consisting of thin hemipelagic and pelagic strata, which accumulated at a very low rate. Their maximum thickness can be recorded in the zone of base discordance. In general, the younger strata gradually become thinner due to a decreasing sediment supply; therefore, the thickness of the transgression stratification system strata decreases toward the paleoslope, the paleobasin, and upsection. The incised paleovalleys are finally buried underneath the transgression stratification system strata.

The high sea-level system crowns the sedimentary sequence. It is characterized by aggradational strata that are replaced by progradational strata with a geometry similar to prograding clinoforms. Strata within the high sea-level system overlie the sequence boundary landward through a base discordance. Seaward, through a baselap, the strata overlie the surface of the transgression stratification system. The high sea-level system comprises the upper and lower clinoforms and subaerial strata. The lower clinoforms are sigmoid in shape, while the upper are oblique. Subaerial strata,

deposited above sea level, are synchronous with the upper clinoforms.

The sedimentary systems of any sedimentary sequence correlate with definite parts of the eustatic sea-level curve. The basin-floor fan of a low sea-level stratification system corresponds to a fast eustatic fall, the slope fan and clinoforms correspond to a late fall or an early rise, the transgression system corresponds to a swift eustatic rise, and the high sea-level system corresponds to the late phase of a eustatic rise, the equilibrium, and the initial phase of the sea-level fall.

Low sea-level stratification systems are characterized by widely distributed, reservoir-quality sands confined to geological bodies of all types. The sands that make up the basin-floor fan mounds are excellent reservoirs with structural and stratigraphic traps for hydrocarbons. Sands carried by turbidity currents thin out or pinch off against paleorises. Stratigraphic traps depend on the top and base seals. The slope fan that often overlies the basin-floor fan is composed of turbidite sands. The stratigraphic traps cannot be traced toward the slope and the positive structures. The sands of a basin-floor fan that are pinched out updip do not trap hydrocarbons. Instead, those hydrocarbons leak upward into the slope fan turbidites. The reservoirs of the lower beds of a slope fan often contain major hydrocarbon reserves. The upper beds contain thin turbidite interbeds without reservoir properties. The sands filling the channels of a slope fan are excellent reservoirs, but they are very difficult to predict ahead of drilling. Their first overlying seal is the back part of the overlying clinoforms containing shallow- and deep-water sands. Shallow-water coastal sands, deltaic sands, and the sands of incised vallies are excellent reservoirs.

Transgression stratification systems contain strata reservoirs confined to the coastal sedimentation zone. The productive strata of many deposits belong to these reservoirs. The overlying clayey sediments of the high sea-level systems form excellent seals. Stratigraphic traps occur in beach ramparts, onlapping sands, and truncations onlapped by clayey sediments.

The condensed (depression) strata of transgression stratification systems are the best oil-source rocks. Continental beds containing coals are formed simultaneously with them.

R.M. Mitchum and J.K. Van Wagoner (1991) identified five levels of the cycles of eustatic sea-level changes: 200, 9 to 10, 1 to 2, 0.1 to 0.2, and 0.01 to 0.02 Ma. The cycles of the fourth and fifth orders can be outlined in areas with a very high rate of sedimentation. The third-order cycle is typical of the above sedimentary sequence. However, according to the data of these researchers, in the sequence of the fourth-order cycle the elements of inner structure described above may also be defined. In this case, the third- and second-order eustatic cycles form a system of sedimentary sequences (megasequences), which, in turn, can be subdivided into stratification systems, making the age hierarchy extremely complicated. Using the data of Vail and his followers, subdivision of the sedimentary cover into sedimentary sequences and stratification systems is

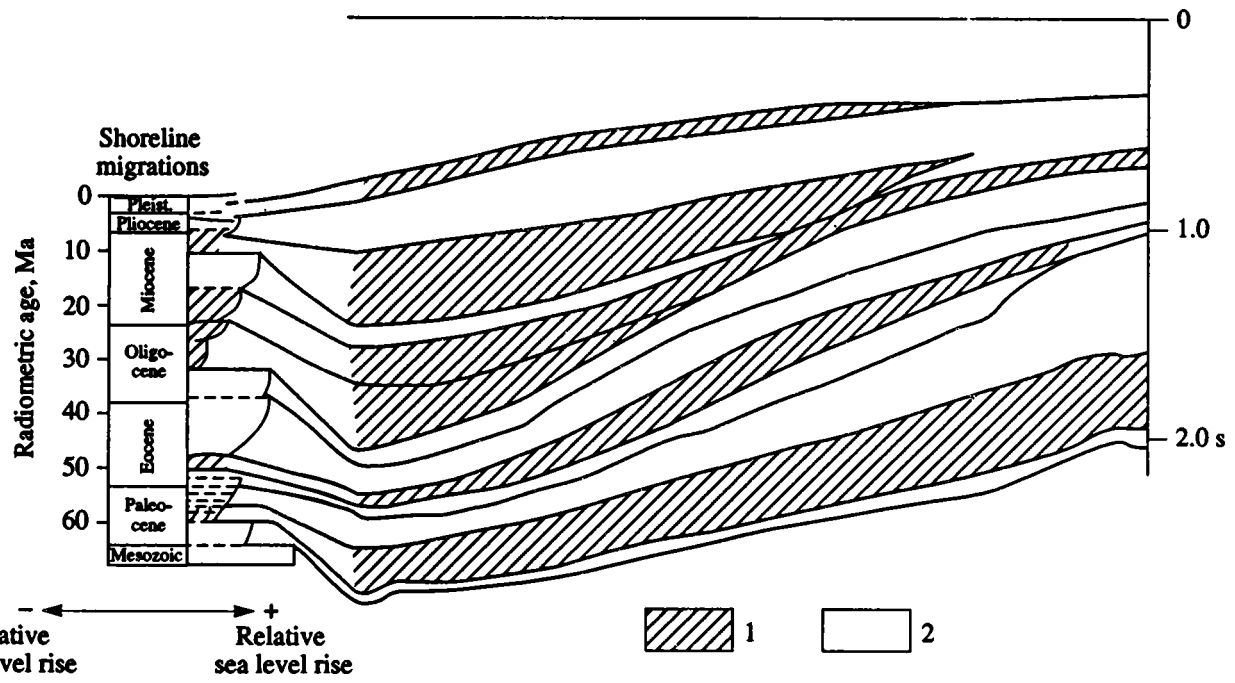

Fig. 5. Part of a seismic time section showing Tertiary deposits in the North Sea (Seismic Stratigraphy..., 1977).
1 – Low sea-level deposits; 2 – high sea-level deposits.

a good way to analyze its age and lithofacies units. The sedimentary sequences are separated by chronostratigraphic boundaries, which give a correct base for correlating and mapping. Interpretation of stratification systems makes it possible to define facies relationships in a sedimentary sequence and to reveal previous sedimentation environments.

SS is based on the global character of unconformities formed by synchronous eustatic falls of the sea level. According to Vail and co-authors, regional tectonic movements are merely superimposed on these falls and complicate the general picture. Thus, a universal procedure for the global correlation of sedimentary cover sections is proposed, which may provide a key framework of stratigraphic units. Their correspondence to a generally adopted stratigraphic scale can be made using biostratigraphical data. In this case, the sedimentary sequence may become the age standard for correcting and controlling stratigraphic correlations, including those obtained using different paleontological groups. Until now, unfortunately, the identification of sedimentary sequences usually involved the use of abstract theoretical suggestions. In most publications, specific sites are not mentioned where a real section could be analyzed based on seismic and well-log data, and where sedimentary sequences and the stratification systems forming them could be identified. This seems to be connected with the confidential status of seismic and well-log data. One publication by Mitchum and Van Wagoner (1991) which is unique in this respect, presents the analysis of the Wilcox and Queen City formations (lower-middle Eocene) of southern Texas in the SS perspective. A detailed correlation of material logged for 295 wells helped to compile geological sections that illustrate the structure of two sedimentary sequences of the third order. The boundaries were dated, through the use of fauna, as 49.5, 48.5, and 46.5 Ma and were correlated with the curve of eustatic cycles (Haq *et al.*, 1988). The transgression and high sea-level stratification systems were identified in the lower sedimentary sequence in sections 108 - 126 km long. In the upper sequence, the same stratification systems and the underlying low sea-level system were recognized, the latter consisting of prograding clinoforms. The sedimentary sequences can be divided into subsystems of the higher fourth-order eustatic cycle. Two condensed levels, confined to the maximum inundation, can be outlined. Subsystem facies are represented by coastal-plain deposits, channel and estuarine sandstones, shallow-water marine sandstones, and shelf and slope shales.

Integration of SS into the theory and practice of stratigraphic investigation requires much additional work; primarily, detailed studies of particular sedimentary basins. In our opinion, the application of SS may be successful in passive sedimentary basins or in their parts that are not involved in active, synsedimentary, tectonic subsidence (Mikhailov and Shlezinger, 1989). In areas with active synsedimentary tectonics, depositional sequences are difficult to identify. It was not simply by chance that Vail (1987) could not interpret a section of a sedimentary cover in the zone of an active growth fault. In a later publication (Vail *et al.*, 1991), using the Gulf of Mexico as an example, he suggested that the elements of the sedimentary sequence considered above are formed in the zone of a growth fault, though this is poorly substantiated by the facts.

In epicontinental basins with intense synsedimentary sinking, as well as in the inner parts of deep basins, the role of eustatic sea-level fluctuations in the formation of their fill is negligible. These fluctuations result only in changes in the facies composition of rocks.

In areas with long-term intense subsidence of the crust, especially in passive continental margins and platforms, some elements of sedimentary sequences can be recognized in certain regions and time intervals (Figs. 2 - 4). For instance, in the Tertiary sediments of the North Sea (Fig. 5) the low sea-level and high sea-level units were defined in a time section (Seismic Stratigraphy..., 1977).

In this country the best sites for the application of SS (Figs. 2 - 4) are the upper Jurassic–lower Cretaceous interval of the west Siberian platform section, the Oligocene–lower Miocene (Maikop) interval of the Scythian platform section, and the upper Devonian–lower Carboniferous interval clinoforms of the Timan–Pechora region. Besides the purely stratigraphical objectives, the suggested studies are of great practical interest, insofar as they promote the identification of the low sea-level paleolevels, when reservoir-quality terrigenous deposits and various nonstructural traps of hydrocarbons were formed. Transgression paleolevels can help to define condensed deposits rich in oil-source matter.

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Evolution of the biosphere

Evolution of Eastern Paratethys in the Late Eocene–Early Miocene

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Abstract – The regional upper Eocene–lower Miocene stratigraphic scale of Eastern Paratethys is developed and the age of the stratigraphic units is correlated with those in the adjoining regions. This is based on malacofauna, microplankton, dinocysts, pollen, and spore assemblages. The following regions are shown in the regional scale: the Black Sea depression, the Ciscaucasian area, the Volga–Don interfluve, the Transcaucasian area (the Akhaltsikhe basin and the Kartli depression), and the Transcaspian area (Mangyshlak, Ustyurt, Kopet Dag). Several regional stages were determined: Beloglinian (upper Eocene), Pshekhian (lower Oligocene), Solenovian (lower Oligocene), Kalmykian (upper Oligocene), Sakaraulian (lower Miocene), and Kotsakhurian (lower Miocene). Lithofacies maps and paleogeographic schemes were compiled for each of the regional stages. All identified stages of the geological evolution of the Eastern Paratethys are supplied with corresponding descriptions of the organic life and climate, the hydrology of the sedimentation basins, and the occurrences of geological resources.

INTRODUCTION

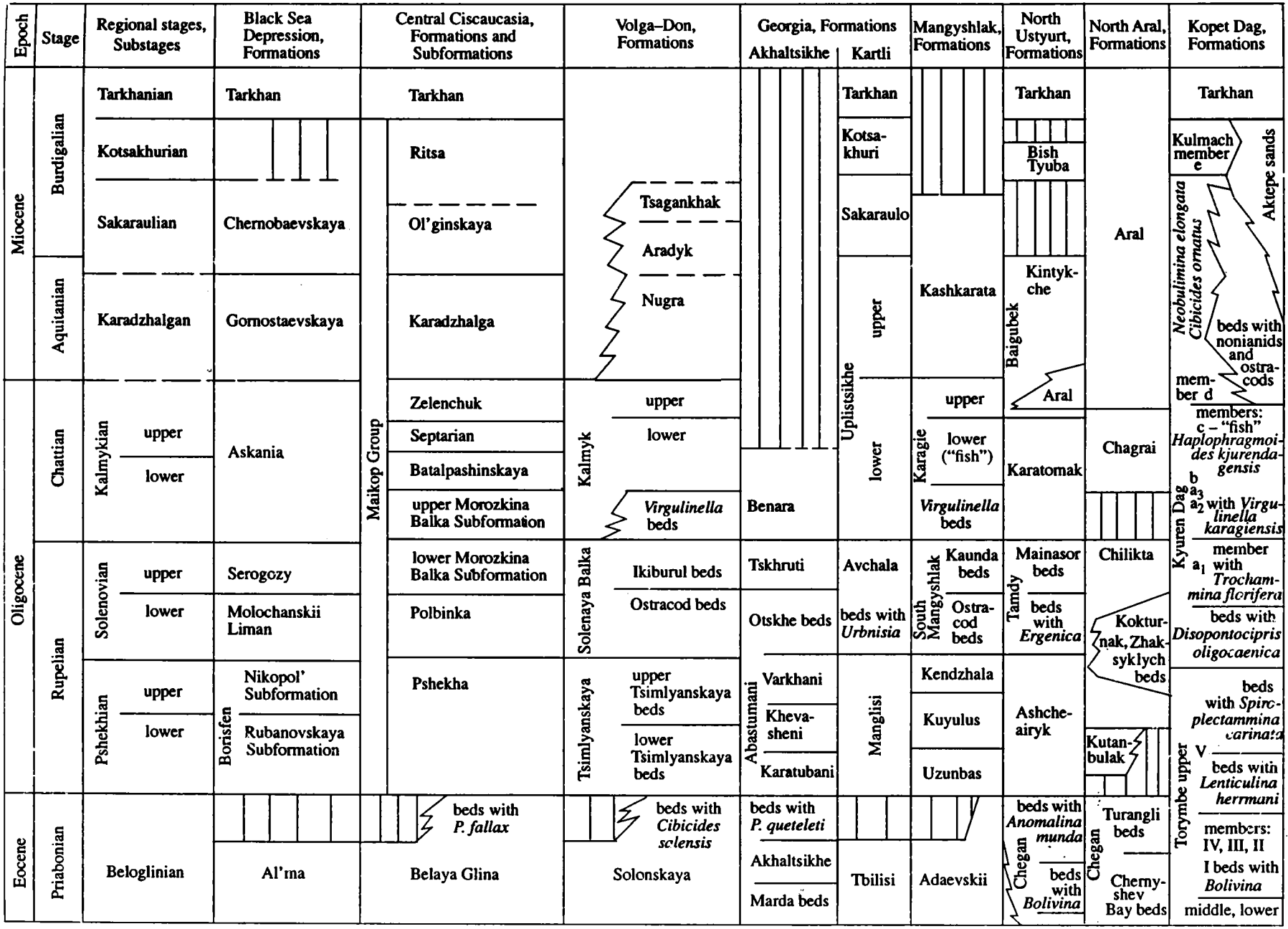
At the Eocene–Oligocene boundary, the eustatic sea-level drop and the uplifting in the Alpine Foldbelt caused the separation of the northern part of the Tethys into two large basins: the Western Paratethys (Baldi, 1980, 1986) and the Eastern Paratethys, which from time to time were directly connected. The isolation of these deep basins, their relatively cold waters, and an abundant inflow of fresh surface and ground water caused stratification and insufficient aeration of their waters and created unfavorable conditions for the life and survival of organisms with carbonate skeletons. As a result, the deposits in these basins contain scanty amounts of fauna, in particular such stratigraphically important groups as plankton foraminifers and nannoplankton. The stratigraphy of these predominantly clayey beds, called the Maikop Group in the Eastern Paratethys, was based primarily on benthic faunal groups and molluscs (Zhizhchenko, 1953; Merklin, 1962, 1974; Voronina and Popov, 1985) and on benthic foraminifers (Ter-Grigor'yants, 1964). Recently, these were supplemented by data on organo-walled phytoplankton. The tracing of lithological units stretching for hundreds of kilometers in the deep-sea part of the basin is crucial for regional correlation (Semenov and Stolyarov, 1970, 1988). At present, the study of the Eastern Paratethys deposits has produced a detailed local stratigraphy of the Maikop Group and a scale of regional stages (Table 1). The units of this scale show the sequence of stages and the transgressive-regressive cyclicity of the basin's evolution.

THE PRINCIPLE LOCALITIES OF THE MAIKOP GROUP AND THE UNDERLYING DEPOSITS

The upper Eocene–lower Oligocene marine deposits are universally observed from the Black Sea to Central Asia. The completeness of the sections, their lithofacies, and the floral compositions vary from one lithostructural zone to another (Table 1).

The Black Sea depression has a complete section of relatively shallow-water deposits of the upper Eocene–lower Miocene, which generally do not crop up on the surface. The faunal assemblages of the more northern type are not abundant but are sufficiently characteristic of and correlative with the Ergeni and Ustyurt assemblages. The rich assemblages of organo-walled

Table 1. Correlation scale of the major upper Eocen–lower Miocene units of the Maikop Group



phytoplankton allows us to identify all the units of the European phytoplankton scale.

Ciscaucasia is a stratotype area used for subdividing the Maikop Group, which fills the large Azov–Kuban' and Terek–Caspian depressions. It is represented mainly by deep-sea sediments. Only the boundary Eocene–Oligocene deposits are well characterized by fauna. Moreover, plankton groups (foraminifers and nannoplankton) alter slightly below the Eocene–Oligocene boundary, e.g., the uppermost part of the Belaya Glina Formation contains assemblages of the *Globigerina tapuriensis* (P18) and *Coccolithus subdistichus* (NP21) zones (Krasheninnikov and Muzylev, 1975). There is an abrupt change in the composition of molluscs, however, which coincides with the boundary between the Belaya Glina and Pshekha formations. In the upper part of the Pshekha Formation the nannoplankton is represented by the assemblages of zones NP22 and NP23 (as determined by A. Nagymarosi). At that boundary, the composition of the organo-walled phytoplankton also changes considerably.

The Volga–Don region is transitional from the deep-sea sections of Ciscaucasia to the more shallow sections on the boundary between the Scythian and Russian platforms. In this area the typical lithological units were identified, and the data on the assemblages of molluscs, palynomorphs, and foraminifers were used to reassess the stratigraphy of the Maikop Group, i.e., to subdivide the Tsimlyanskaya Formation into two members, to identify beds containing ostracods and the Ikiburul beds in the Solenaya Balka Formation, and to distinguish the Kalmyk Formation with its two members (Semenov and Stolyarov, 1988; Voronina *et al.*, 1988). The correlation of these units with sections of Ciscaucasia, Mangyshlak, and Ustyurt, on the one hand, and those of Ukraine, on the other, is sufficiently reliable.

Transcaucasia

The sections of the Akhaltsikhe and Kartli depressions in Georgia are unique for their completeness and faunal characteristics; moreover, they supercede one another.

The Akhaltsikhe depression is filled with middle Eocene–lower Oligocene marine sediments containing diverse fauna. The molluscs, foraminifers, and nummulites of the upper Eocene are rich, warm-water communities of predominantly Tethys origin, and they are strikingly different from the more northern assemblages. The lower Oligocene molluscs (about 90 species) are also indicative of a relatively warm-water environment (genera *Atrina, Megaxinus, Lutraria, Arcopagia,* etc.). However, the abundant species in common with the more northern lower Oligocene assemblages of the Eastern Paratethys are dominant. The brackish-water molluscs in the Otskhe beds and in the Tskhruty Formation (Popov and Titova, 1982) are likewise diverse and are represented by the same endemic genera as found throughout the Paratethys, but with a greater variety of cardiids and corbulids.

The Kartli depression is a large syncline filled with marine sediments of the Eocene–Miocene age, within which the transitions can be traced from deep- to shallow-water, coarse clastic sections containing a richer variety of fauna (Voronina *et al.*, 1991). In this region, the most interesting assemblages of molluscs are those from the Uplistsikhe Formation (more than 30 species) and from the Sakaraulo Creek (about 100 species) and Mt. Kotsakhuris (20 species) sections, which are unique in Eastern Paratethys and represent the southern type of fauna.

Transcaspian

Three major subregions are identified within the Transcaspian area: Mangyshlak, Ustyurt, and Kopet Dag.

Mangyshlak. The numerous bore holes and outcrops in this area provided data for a detailed study of the facies transition from the coastal (littoral) deposits with rich faunal assemblages, across the outer shelf with manganese ore formations, to the pseudobathyal zone in the southwestern part of Mangyshlak with deep-sea type sediments. Over the short, about 100 km, distance of this transition, not only the lithofacies but also the thickness of the layers in the separate Maikop units vary greatly.

North Ustyurt. The outcrops of the Ustyurt scarps reveal relatively shallow-water coastal deposits in the late Eocene and Oligocene basins with extremely rich assemblages of benthic fauna of the northern type (Mironova *et al.*, 1958; Merklin, 1974; etc.) and of organo-walled phytoplankton. Molluscs, ostracods, foraminifers, and cysts of dinoflagellates from these deposits show good correlation with the assemblages of northern Germany and Belgium.

Kopet Dag. The sections of western Kopet Dag are complete with a distinct sequence of microfaunal zones (Table 1) but poor microfaunal assemblages. In the east, the deposits are of a more shallow-water type, less complete, and contain the remains of molluscs. The characteristic fauna of the East Caspian part of the basin belongs mostly to the northern type with individual elements of a more warm-water variety.

THE REGIONAL STRATIGRAPHIC SCALE OF THE UPPER EOCENE–LOWER MIOCENE IN THE EASTERN PARATETHYS

The stratigraphic boundaries in the Oligocene–lower Miocene deposits of the different regions of the Eastern Paratethys are established at the levels recording the crucial changes in the conditions of sediment accumulation and in the fauna of the basin (Table 1). These levels form a sound basis for the compilation of the regional stratigraphic scale of the Eastern Paratethys. Earlier,

similar attempts were made (Vyalov, 1940, 1964; Zhizhchenko, 1953), but at that time the data sample of the Maikop deposits was inadequate to base the regional units on important geological events, some of which could have global significance. The horizons, later identified by R.L. Merklin (1962), were determined from the incomplete and atypical sections of northern Ustyurt. We consider G.I. Semenov's attempt (1968) to be the most successful subdivision of the Maikop strata into four transgression-regression cycles. Three of them are acknowledged as regional stratigraphic units of the same volume.

The Beloglinian regional stage (Korotkov, 1935; Stratigraficheskii Slovar', 1937). The stratotype of this stage was identified in the Kuban' River section upstream of Cherkessk. The regional stage was observed in the typical facies of white carbonate clays and marls from the South Aral region to southern Ukraine and further to central Europe. The beds with *Globigerapsis tropicalis* and *Globigerina corpulenta* contain rich assemblages of foraminifers (zones P15 and P16) and nannoplankton of the joined *Discoaster barbadiensis* Zone (NP18 - NP20) (Table 2). In layers with *Bolivina antegressa*, the composition of planktonic foraminifers is impoverished, and the species are more typical of the Oligocene. Individual lithostructural zones, even in the open, relatively deep-water parts of the basin show traces of shallowing with the reworking of deposits and the appearance of benthic fauna (molluscs *Propeamussium fallax, Pychodonte queteleti* and others). This locality also contains plankton foraminifers of the *Globigerina tapuriensis* Zone (P18) and the nannoplankton of the *Coccolithus subdistichus* Zone (NP21) of the Oligocene base (Krasheninnikov and Muzylev, 1975). In the shallower part, the beds with *Bolivina antegressa* are substituted in their upper portion by terrigenous rocks (beds with *Asterigerina lucida* and *Cibicides salensis*, beds with *Anomalina munda* of the Chegan Formation, and their analogs). This sequence incorporates the microfauna of the transitional Belaya Glina–Khadum beds and the warm-water molluscs similar to the Eocene assemblages. At that boundary, cardinal changes took place in the organo-walled phytoplankton assemblages. The typical Eocene *Areosphaeridium arcuatum* and *A. diktyoplokus* disappear, as do the species of genera *Cordosphaeridium, Spiniferites,* and, among green algae, *Tytthodiscus beloglinensis.* A number of species of genus *Phthanoperidinium* appeared, which were well adapted to habitats with changing salinity and lower temperatures.

The Pshekhian stage stratotype (Korotkov, 1935) is located on the Pshekha River in western Ciscaucasia; the hypostratotype outcrops on the left-hand bank of the Kuban' River near Cherkessk. In the relatively deep-sea zone, this stage is usually represented by dark, irregularly carbonate, silty clays. The base of the horizon contains the planktonic foraminifers of the *Globigerina tapuriensis* Zone (P18), the nannoplankton of the *Coccolithus subdistichus* Zone (NP21); the middle part of the horizon includes the nannoplankton of the *Helicopontosphaera reticulata* Zone (NP22), and the upper part of the *Sphenolithus predistensis* Zone (NP23, after Nagymarosy). According to N.G. Muzylev, this zone is transitional between NP22 and NP23.

Two units can be identified in the shallower zone: the lower clayey part with foraminifers of the *Lenticulina herrmanni* Zone, which contains the typical mollusc assemblage with *Yoldiella chadumica, Pterolucina batalpaschinica, Palliolum simile*; and the upper part usually silty, with occasional sand interbeds. The benthic foraminifers of the upper part are represented by the assemblage with *Spiroplectammina carinata oligocenica.* The molluscs include larger, more shallow-water forms: *Nemocardium excomatulum, Arctica rotundata,* and characteristic *Pecten stettenensis.* In different localities these units are not strictly coeval, but they are widely distributed and can be adopted as the upper and lower substages of the Pshekhian Stage.

The Pshekhian Stage includes the three phytoplankton zones of the boreal dinocysts scale of West Europe, i.e., *Phthanoperidinium amoenum, Wetzeliella symmetrica,* and *W. gochtii.* The first zone corresponds mainly to the lower substage. In the assemblage of palinomorphs of higher plants, Taxodiaceae and the pollen of Pinaceae are the most important, particularly in the basal beds of the Pshekhian Stage, which correlate with the cooling at the Eocene–Oligocene boundary.

A discovery vital for the interpretation of paleogeographical environment and climatic conditions was made in the Pshekhian beds of the Belaya River, north of the Abadzekhskaya railway station, by A.F. Bannikov of the Paleontological Institute, Moscow. It was here that M.A. Akhmet'ev determined the fragments of monocotyledons to be *Taxodium dubium* (Sternb.) Heer, *Cinnamomum* sp., *Myrtophyllum* sp., *Apocynophyllum helveticum* Heer, and *Dryophyllum furcinerve* (Rossm.) Schmalh. These plants indicate both a subtropical climate in a coastal zone and the presence of island archipelagoes in the area of the Greater Caucasus.

The magnetostratigraphy provides a good control for the Pshekhian sequence (Molostovskii, 1986; Kunaev, 1990), because it belongs almost entirely to the zone of reversed polarity, the only exception being the Eocene boundary, which has three normally magnetized zones. This observation is confirmed by the study of the Kuban' and Ashcheairyk sections in north Ustyurt, and of the Kutanbulak Formation sections in Perovskii Bay in the northern Aral area.

The Solenovian Stage stratotype (Popov, 1959) is the bore hole near the Tsimlyanskii Settlement on the Don River; the parastratotype is located on the left bank of the Kuban' River facing Cherkessk. It is divided into two substages (Voronina and Popov, 1984). The lower substage, observed in the Volga–Don region corresponds to the ostracod layers and is represented by a marker layer of light-colored marls or carbonate clays

Table 2. Paleontology and paleoecology of the regional stages of the Eastern Paratethys

Epoch	Stage	Regional stages (horizons)	Foraminifers	Bivalve molluscs	Dinocysts scale: zone	Dinocysts scale: subzone, beds	Nearshore salinity (10‰ 20‰ 30‰)	Transgression / Regression
Miocene	Burdigalian	Tarkhanian		*Crassostrea gigas* *Lentipecten denudatus*				
		Kotsakhurian	*Saccammina zuramakensis*	*Rzehakia dubiosa* *Eoprosodacna*				
		Sakaraulian	*Neobulimina elongata*	*Palliolum simile* *Thyasira tautilus* *Fragum semirugosum, Acanthocardia grande* *Pholadomya alpina*				
	Aquitanian	Karadzhalgan	*Cibicides ornatus* *Elphidium onerosum* *Porosononion dendriticus*	*Palliolum incomparabile* *Callista lilacinoides* *Venus multilamella*				
Oligocene	Chattian	Kalmykian upper		*Cerastoderma prigorovskii* *Corbula helmerseni*		*Deflandrea spinulosa* *Homotriblium floripes*		
		Kalmykian lower	*Spiroplectammina terekensis, Sphaeroidina variabilis* *Virgulinella* ex gr. *pertusa* *V. karagiensis*	*Plagiocardium abundans* *Chlamys bifida*	*Chiropteridium partispinatum*	*Rhombodinium draco*		
	Rupelian	Solenovian upper	*Trochammina florifera*	*Janschinella, Cyrtodaria, Nucula*				
		Solenovian lower	Ostracods *Disopontocypris oligocaenica*	*Ergenica, Urbnisia, Korobkoviella*	*Wetzeliella gochtii*			
		Pshekhian upper	*Spiroplectammina carinata oligocenica*	*Flabellipecten stettenensis*	*Wetzeliella symmetrica*			
		Pshekhian lower	*Lenticulina herrmanni*	*Joldiella chadumica* *Pterolucina batalpaschinica*	*Phthanoperidinium amoenum*			
Eocene	Priabonian	Beloglinian	*Cibicides salensis*	*Nucula aralensis* *Aralocardia* / *P. fallax*	*Wetzeliella clathrata angulosa*			
			Bolivina antegressa	*Palliolum mayeri* *Pycnodonte queteleti*				
			Globigerapsis tropicalis *Globigerina corpulenta*					

with local interbeds of carbonate-free varieties. Almost everywhere this substage contains diverse and abundant fauna with typicaly large ostracods *Disopontocypris oligocaenica*, etc., a peculiar nannoplankton assemblage with *Reticulofenestra ornata*, and other species placing this assemblage in the lower part of the *Sphenolithus predistentus* Zone (NP23). The molluscs here are represented by endemic brackish-water forms: *Ergenica, Urbnisia, Merklinicardium, Korobkoviella,* and *Janschinella.*

The dinocysts belonging to the *Wetzeliella gochtii* Zone are represented by an impoverished assemblage containing *Gerdiocysta*, which is accompanied by the *Cyclonephelium–Oligosphaeridium–Operculodinium* group, and by the genera *Distatodinium* and *Hystrichokolpoma.* A similar composition is recorded both in the lower subhorizon (ostracod beds) and in the overlying deposits. The solitary finds of phytoplankton, typical of the basins connected with the Norwegian-Greenland basin, indicate constant, though partially obstructed, ties with the Northern Atlantic through either the Mainz basin or the Masuria–Mazovia threshold, as suggested by A. Rusu (1988).

The upper substage (the Ikiburul beds of the Volga–Don region) is composed of carbonate-free, dark or light-colored clays with poor fossil remains, which are sometimes lithologycally similar to the overlying clays. In the marginal parts of the basin, it contains brackish-water fauna. However, some marine molluscs and impoverished assemblages of agglutinated foraminifers (*Paragaudryina tripartita* and *Trochammina florifera*) are also present. The palino-assemblages of the Solenovian Horizon are characterized by a higher content of pines and beeches and by diverse thermophilic forms. Paleomagnetic data shows that the larger part of the Solenovian regional stage, and likewise the Pshekhian Stage, belong to the R-zone with two N-events in its second half.

The Kalmykian regional stage is identified for the first time in this paper. Its stratotype is represented by the bore hole section, which was drilled 50 km south-southeast of Elista. Semenov and Stolyarov (1988) previously described the stratotype of the Kalmyk Formation in their paper. At the base of the regional stage, in a relatively deep-sea zone, a bed of dark carbonate clays occurs, which contains *Virgulinella* ex gr. *pertusa, V. karagiensis,* and the nannoplankton of the *Sphenolithus distentus* Zone NP24. In areas where the "*Virgulinella* bed" is absent, the Kalmykian Stage begins with dark, carbonate-free clays containing scattered fish remains, the so-called "fish beds" (Semenov and Stolyarov, 1970). In the more shallow-water sections there are no fish remains, and the beds on that level contain benthic foraminifers of the *Spiroplectammina terekensis* Zone with abundant *Uvigerinella californica* and the marine molluscs *Chlamys bifida, Nuculana gracilis, Plagiocardium abundans,* etc. The *Virgulinella* and fish beds (together with their more shallow-water analogs) compose the lower substage. The upper substage is connected with the lower one by a gradual transition, and is composed of either silty or sandy rocks with impoverished assemblages of benthic foraminifers and molluscs, containing the typical *Cerastoderma prigorovskii* and *Corbula helmerseni.* Within this stage, the assemblage of the organo-walled phytoplankton belongs to two dinocyst zones having a wide regional distribution: the lower, *Chiropteridium partispinatum–Rhombodinium draco* Zone, and the upper *Deflandrea spinulosa–Homotriblium tenuispinulosum* Zone. This part of the Maikop Group includes 5 to 6 large zones of normal polarity; their distinct grouping is observed in a number of interrelated sections. The uppermost Oligocene beds are characterized by a predominant R-polarity (Molostovskii, 1986).

The overlying deposits are the most monotonous and poor in fauna, and so their subdivision and correlation are very difficult.

The Karadzhalgan regional stage (Prokopov, 1937). The stratotype is located on the Karadzhalga River in the western Ciscaucasia, and the hypostratotype is located on the Kuban' River downstream of Cherkessk. This stage consists mainly of deep-water, clayey deposits. In certain localities containing shallow-water types of sections, a change can be observed from the impoverished euryhaline mollusc complex of the upper Kalmykian Stage to the richer polyhaline complex with warm-water genera and species of Tethys origin, which are also found in the western Paratethys: *Callista lilacinoides, Palliolum incomporabile, Isognomon,* etc. Tentatively, we correlate this level of faunal change with the base of the Karadzhalgan regional stage. The sudden changes in the composition of molluscs at this level were not accompanied by changes in other groups; foraminifers are represented by the same poor assemblage containing *Cibicides ornatus* and *Ephidium onerosum;* the dinoflagellates belong to the *Deflandrea spinulosa–Homotriblium tenuispinulosum* Zone, which is typical of the underlying rocks. In the pollen assemblages of the upper part of the Karadzhalga Formation, the pollen of gymnosperms dominate (up to 90 - 95%), probably because of the active uplifts in the area of the Greater Caucasus.

The Sakaraulian regional stage (Davitashvili, 1933) has its stratotype at Sakaraulo Creek facing the town of Kaspi, while the hypostratotype is located at Naderbazevi Creek near the Metekhi railway station (Voronina *et al.,* 1991). In the stratotype region of central Georgia, this stage is represented by coarse-grained granule sandstones in the lower part and by clayey sandstones and siltstones in the upper part. The stage contains only the remains of the thermophilic and shallow-water mollusc assemblage. The deposits of the Ol'ginskaya Formation, identified also in the sections of the Maikop Group by the coarser sediments and by the appearance of a new assemblage of foraminifers with *Neobulimina elongata,* are considered analogous to the Sakaraulian

Stage in Ciscaucasia. By tracing the above microfaunal assemblage, the analogs of the Ol'ginskaya Formation (Sakaraulian Stage?) are found in the Black Sea region, in the Crimea, and in the South Aral area.

The Kotsakhurian regional stage (Davitashvili, 1933). The stratotype is located at the town of Kaspi, while the hypostratotype is located at the Naderbazevi Creek near the Metekhi railway station (Popov and Voronina, 1988). In the Kartli depression, this stage is represented by silty sandstones and clayey siltstones with sandstone interbeds. The fauna of the stratotype region is unique. However, correlation with the Kotsakhurian deposits is reliable, based on indications of lower salinity, e.g., the impoverished assemblage of foraminifers with *Saccammina zuramakensis* and brackish-water molluscs with *Rzehakia dubiosa*. The Tarkhan, or Chokrak, deposits in the marginal parts of the basin overlie the Kotsakhurian beds with obvious unconformity and considerable hiatus.

CORRELATION AND AGE OF REGIONAL STRATIGRAPHIC UNITS

The lower two microfaunal zones of the Beloglinian regional stage, *Globigerapsis tropicalis, Globigerina corpulenta,* and the *Bolivina antegressa* Zone, correspond according to plankton foraminifers and nannoplankton to the Priabonian in the stratotype. Less definite is the age of the upper *Cibicides salensis* Zone, the definition of which is directly associated with the determination of the Eocene–Oligocene boundary.

In the Boreal region, this boundary, assumed to be the boundary between the Priabonian and Rupelian s.l. analogs, is not identified and, therefore, in the course of the works in the framework of the IGCP project 174 *Geological events at the Eocene–Oligocene boundary* it was tentatively placed at the base of the *Globigerina tapuriensis* Zone (P18, Table 3) inside the zone NP21. The data on plankton foraminifers place the Eocene–Oligocene boundary in the upper Beloglinian, whereas in central Europe it lies in the lithologically monotonous beds of the Buda (Hungary) and Brebi (Romania) marls. On the vast territory of Eurasia, the changes in sedimentation and water circulation, the peak of regression that determined the migration of vertebrates, the appearance of the boreal and endemic groups of marine fauna, i.e., all the most important paleogeographical and evolutionary changes, occurred directly after this boundary. We consider it necessary to define more clearly the time and scale of these changes, but meanwhile we are assuming that the Eocene–Oligocene boundary lies at the base of the Maikop Group, which approximately corresponds to the base of the *Helicopontosphaera reticulata* Zone (NP22) (Resolutions of the Paleogene Commission of the Interdepartmental Stratigraphic Committee, adopted in 1959 and 1963).

Through the planktonic and benthic fauna, the Pshekhian regional stage has a reliable correlation with the lower Rupelian s.l., which includes the upper part of the Tongrian (Table 3) and is also characterized by the molluscs *Nucula comta, Pecten stettenensis, Astarte plicata, Callista beirichi,* etc. Through the nannoplankton and cysts of dinoflagellates, and by its position in the sections, the Solenovian Stage correlates with the Rupelian s.s.

The lower part of the Kalmykian Stage likewise reliably correlates with the lower Chattian through the finds of nannoplankton (zone NP24 in the *Virgulinella* beds), dinoflagellates, and molluscs, including the zonal forms of Chattian A (*Chlamys bifida*). Defining the position of the Paleogene–Neogene boundary and the age of the Karadzhalgan regional stage is more complicated. The discoveries of dinoflagellates of the genera *Wetzeliella* and *Rhombodinium*, and the presence of *Deflandrea spinulosa* in the Karadzhalga and Gornostaevka formations and in the Sivash beds correlate these deposits with the Chattian (Andreeva–Grigorovich, 1980). Apparently, the renovation of the mollusc assemblages, the appearance of the upper Egerian index species, and their enrichment with polyhaline warm-water groups of the Tethys origin characteristic of the Neogene occurred, after the Karadzhalgan. Therefore, we tentatively correlate the Karadzhalgan Stage with the Miocene part of the Egerian in central Europe and with the Aquitanian, attributing the entire Kalmykian Stage to the Chattian. However, we consider the Sakaraulo and Ol'ginskaya formations to be analogs of the Eggenburgian and Burdigalian. The Kotsakhurian Stage reliably correlates with the upper Ottnangian in central Europe by the presence of *Rzehakia dubiosa, Cerastoderma, Limnopappia,* and other brackish-water forms.

THE BASINS OF EASTERN PARATETHYS: PECULIARITIES OF SEDIMENTATION AND BIOTA

The separation of the Eastern Paratethys as an intracontinental basin with specific conditions of deposition and its own geological history and fauna coincided both with the change of predominantly carbonate sediments into terrigenous ones, and with the beginning of the formation of the Maikop Group. Many features of this basin, however, appeared earlier, e.g., the bathymetry and the paleogeographic ties.

The Beloglinian Basin

Paleogeography and sedimentation. The late Eocene basin was the largest sea basin in the Cenozoic. It covered the vast area from the Carpathians to Central Asia and western Siberia (Fig. 1).

The uplifts of the Alpine Foldbelt separated the basin from Tethys. However, with a high eustatic sea-level, the basin was connected with Tethys by straits in the area of the Lesser Caucasus and in Asia Minor. A free connection with the World Ocean also existed in the west through the Norwegian-Greenland basin. These long-standing and stable ties were recorded in

Table 3. Correlation scheme of upper Eocene–lower Miocene stratigraphic units of Europe

Age, Ma	Epoch	Stage	Plankton zones: Foraminifers (Blow, 1969, 1979)	Plankton zones: Nannoplankton (Martini, 1971)	England	Paris basin	Belgium	North Germany	Mainz basin	Hungary (Czechia, Slovakia, Austria)	Roumania (Transylvania)	Eastern Paratethys
15	Miocene	Langhian	N8 *Globigerinoides sicanus*	NN5 *Sphenolithus heteromorphus*			Antwerp sands			Badenian	Badenian	Chokrakian
				NN4 *Helicopontosphaera ampliaperta*						Carpathian	Hida beds	Tarkhanian
		Burdigalian	N7 *Globigerinoides trilobus*				Edeghem	Hemmoor				Kotsakhurian
20			N6 *Globigerinita dissimilis*	NN3 *Sphenolithus belemnos*						Ottnangian	Chechis marls	Sakaraulian
		Aquitanian	N5 *Globoquadrina dehiscens*	NN2 *Discoaster druggi*						Eggenburgian	Corus	
			N4 *Globorotalia kugleri*	NN1 *Triquetrorhabdulus carinatus*				Vierland				
			P22 *Globigerina ciperoensis*	NP25 *Sphenolithus ciperoensis*				C – upper Dohberg sands	Cerithian beds	Egerian	Vima clays	Karadzhalgan
25		Chattian	P21 *Globorotalia opima opima*	NP24 *Sphenolithus distentus*				Chattian				
			P20 *Globigerina ampliapertura*			Boos Limestones	Voort sands	B – lower Dohberg	Fish shales?			Kalmykian
30	Oligocene							A – Kassel sands		Kiscel clays	Buzas sands	
		Rupelian	P19 *Globigerina sellii*	NP23 *Sphenolithus praedistentus*	Hamstead	Stampian s. l. Stampian s. s.	Rupelian Boom clays	Septarian clay	Foraminiferal marl	clays upper	Ileanda clays	Solenovian
							Clays with *N.comta*		lower sands	Beds with *Cardium lipoldi*	Bizusa marls	
			P18 *Globigerina tapuriensis*	NP22 *Helicopontosphaera reticulata*		Sannois	Berg sands Vieux–Jones sands upper Tongrian Henis sands		Pechelbronn beds	Tard lower	Mera beds Hoia limestones	Pshekhian
35	Eocene	Priabonian	P17 *Globorotalia centralis*	NP21 *Coccolithus subdistichus*								
				NP20 *Sph. pseudoradians*	Bembridge	Ludian Montmartre marls and gypsum	lower Tongrian Glimmertingen sands	Lattorfian sands		Buda marls	Brebi marls	Beloglinian
			P16 *G. cocoaensis*	NP19 *Isth. recurvus*								
			P15 *Globigerapsis semiinvoluta*	NP18 *Chiasmolithus oamaruensis*	Osborne Headon	Marls with *Pholadomya ludensis*		Schonenwalder Formation		Bryozoan marl	Cluj limestones	
40					Barton	Marinesian sands	Assche sands					Kumanian

the marine deposits of that age and in the distribution of fauna. The salinity of the basin was close to normal.

The basic facies paleogeographical elements of the Eastern Paratethys basin appeared at the Cretaceous–Paleogene boundary (Shutskaya, 1970; Tugolesov *et al.*, 1985; Stolyarov, 1991) and for a long time followed the inherited pattern. The most striking feature of the basin was its more or less obvious morphological division into areas with very different sedimentation in the shallow-water shelf (marginal) seas and in the vast open deep-sea parts of the basin. The latter covered the areas of the Black Sea, the Sea of Azov, Ciscaucasia, the Caspian Sea, the Transcaspian depressions of South Mangyshlak, Kopet Dag, and eastern Transcaucasia.

The position of the deep-sea depressions with entirely carbonate sedimentation and of the shelf seas with mainly terrigenous (sandy–clayey) and siliceous sediments was controlled by the positive paleogeographic (morphological) elements that divided the areas with different types of deposits. In the northern part of the epicontinental basin, the sublatitudinal island system formed a paleogeographic barrier within the Ukrainian crystalline massif and Donets Basin, which bordered the extensive shelf sea of the Dnieper–Donets depression. In western Ciscaucasia (Stolyarov, 1991), an extended arc-like Ergeni–Stavropol'–Tomashevskaya band of low islands and shallows marking the outer shelf existed. In the south, the island groups in the eastern part of the Greater Caucasus and in the Georgian block divided the deepest sea areas of Ciscaucasia and the Rioni depression from the inner shallow seas of Transcaucasia with very thick, terrigenous and volcanic sediments (up to 1200 m). The morphology of the northeastern part of the basin (the Caspian depression and the Aral area) was more leveled and the transition from carbonate to terrigenous sedimentation was gradual.

Late Eocene sedimentation and tectonic subsidence occurred simultaneously. By the end of the Eocene the basin became shallower, causing a change in the foraminifers, with a substitution of planktonic assemblages by predominantly benthic organisms of the *Bolivina* Zone. This zone was not recognized in the sequence of the deepest part of the basin in eastern Ciscaucasia, which was not compensated by sedimentation and had no "regressive cycle".

Organic life. Molluscs. The deep-sea facies were barren of molluscs, but during the regression at the end of Beloglinian time, the assemblage of *Propeamussium fallas* was widely distributed on carbonate oozes at depths of about 100 - 200 m. The richest shallow-water assemblages on the sandy silt substratum are strikingly different in each of the three regions:

(1) on the slopes of the Ukrainian shield, mainly in the Mandrikovo Complex, the bivalve molluscs were represented by no less than 100 species, 60 genera, and 31 families, and included the tropical genera *Fragum* and *Fimbria;*

(2) the Transcaucasian part of the basin was even richer in bivalve molluscs (no less than 170 species, 77 genera, and 36 families); these molluscs were more thermophilic and were represented by the Tethys genera and species;

(3) in the Transcaspian region, the initial composition of fauna was much less diverse, but at the expence of thorough study about 120 species, 67 genera, and 38 families of bivalve molluscs were recorded. The more thermophilic genera, such as *Arca, Spondylus,* and *Chama* are missing, but there are many endemic species (about one third) and rather rich assemblages lacking individual domination. On sandy substratum, the typical organisms were *Callista latilamella, Nemocardium rasmyslovae, Pelecyora tenuis, Cardites, Atrina,* and *Aralocardia.*

Phytoplankton. The principal components of the organo-walled phytoplankton assemblage (*Wetzeliella clathrata angulosa* Zone) were species of the genera *Deflandrea, Cordosphaeridium, Areosphaeridium, Microdinium,* and *Dinopterigium,* and also the genera of the *Wetzeliella* group: *Charlesdowniea, Wetzeliella, Kisselovia,* and *Rhombodinium.* The dominant taxa among the green algae and acritarchs were those typical of warm-water basins with normal salinity, e.g., *Crassosphaera, Tytthodiscus,* and *Lanjaria.*

Flora and vegetation. The basic zonal vegetation in early Beloglinian time was the oak-laurel forest, which mixed with Pinaceae and Taxodiaceae in the mountains. The spectrum of thermophilic broad-leaved plants included Araliaceae, Myricaceae, Moraceae, Hamamelidaceae, and Anacardiaceae. On the individual islands of the archipelagoes in the more southern regions, the coastal vegetation consisted mostly of palms and thermophilic ferns (*Blechnum,* etc.). The fern *Azolla vera* was found buried in the early Beloglinian deposits in the eastern part of the basin. By the end of Beloglinian time, as the climate became cooler, the proportion of angiosperms in the vegetation noticeably increased.

Climate. In the Transcaucasian part of the basin, at the beginning of Beloglinian time, the benthic marine fauna inhabited very warm waters similar to those in present-day tropical and subtropical areas. At a later stage, the composition of molluscs implies a gradual cooling of the waters, probably caused by the obstruction of connections with the Tethys and by global cooling in general. The latter event is also confirmed by paleofloral data (Akhmet'ev and Zaporozhets, 1992).

In the western part of the basin, on the slopes of the Ukrainian shield, at the beginning of Beloglinian time, the water was also very warm and inhabited by tropical and subtropical fauna; there were fewer tropical species of molluscs, however, than in Transcaucasia. The flora in this area was close to humid subtropical and the spore-pollen spectra belonged to forests with the angiosperms represented mainly by evergreens and laurels (Zaklinskaya, 1958). In the more easterly regions of Europe and in Transcaucasia, the spectra are indica-

tive of a periodic aridization of the climate. Individual phases indicative of a drier climate are recorded in the upper Eocene portion of the Landzhar section in Armenia (Zaporozhets, 1989). The benthic fauna of molluscs in the Transcaspian part of the basin was the coldest water of the Eocene faunas in western Eurasia, i.e., species of the tropical genera were missing, the taxonomic diversity was low, and species of the subtropical genera constituted about one third to one half of the species in the assemblages.

According to palynological data, at the time of the Chegan deposit accumulation, the Aral and Turgai areas were covered with subtropical forests, where evergreen sclerophyllous Myricaceae, Lauraceae, and Fagaceae were dominant. In southern Central Asia, at the time of the Sumsar deposit accumulation, broad-leaved forests were dominant. They also contained an appreciable number of tropical plants (palms, Moraceae, *Liquidambar,* and *Platanus*) and ferns (Pulatova, 1983). A considerable amount of *Ephedra* grew in the open aridic areas.

The Pshekhian Basin.

Paleogeography and sedimentation. After regression at the beginning of Pshekhian time there was a new ingression of the waters, and in the beginning of the Oligocene the basin stretched latitudinally for more than 4500 km and longitudinally for 1500 km (Fig. 2). The boundaries of the early Pshekhian basin, like those of the Beloglinian basin, extended beyond the modern area of their deposits. This is implied not only by the sparse shallow-water facies, but also by the early Oligocene fauna discovered far beyond its usual distribution in sediments previously considered to be continental, e.g., in the Kurgan beds in the southern part of West Siberia. And yet, the Pshekhian basin never reached the size of the former late Eocene basin; it had a restricted water exchange with the World Ocean (as compared to that of the Beloglinian basin) and with Tethys. The main connection with the open sea was that in the west with the Norwegian-Greenland basin. The faunal exchange also occurred through that channel.

The general paleogeographic conditions of the Pshekhian basin were inherited from the late Eocene. The intense tectonic movements at the Eocene–Oligocene boundary caused essential transformations in its water area and had an influence on the paleogeography and dynamics of sedimentation. The primary difference between the early Oligocene and the late Eocene basins is a significant increase in the depth of the shelf depressions, which reached 1000 m and deeper, whereas the hypsometric position of the major positive paleogeographic elements, separating the seas with different types of sedimentation, remained sufficiently stable. This situation caused a stronger morphological differentiation in the Eastern Paratethys and the appearance of major benches in the edge areas of the shelf (continental slopes). It also distinctly revealed one of the principle paleogeographic lineaments of the basin, i.e., an elongated, intermittent system of low islands and shallows stretching from the Ukrainian shield and Donets Basin across Karpinskii Ridge and the mountains of Mangyshlak to Central Ustyurt. This system separated the shelf seas of the Dnieper–Donets depression, the Volga–Don area, the Caspian depression, and the Aral area from the off-shelf, deep-sea basins of the Black Sea, South Ukraine (Tugolesov *et al.*, 1985), eastern Ciscaucasia, and the central Caspian Basin. The characteristic feature of these basins is the uncompensating regime of sedimentation with deep-water calcareous-clayey sediments only 10 - 30 m thick. The Greater Caucasus–Transcaucasia island archipelago and the islands of the Crimea mountain massif occupied isolated positions in the south. The subsidence of the Black Sea depression is testified to by an area of widely developed Oligocene deposits, as compared with those of the Paleocene–Eocene, and by their transgressive occurrence on the slopes of large synsedimentary uplifts within the depression (Andrusov Rise, Shatskii Rise, etc.) (Tugolesov *et al.*, 1985).

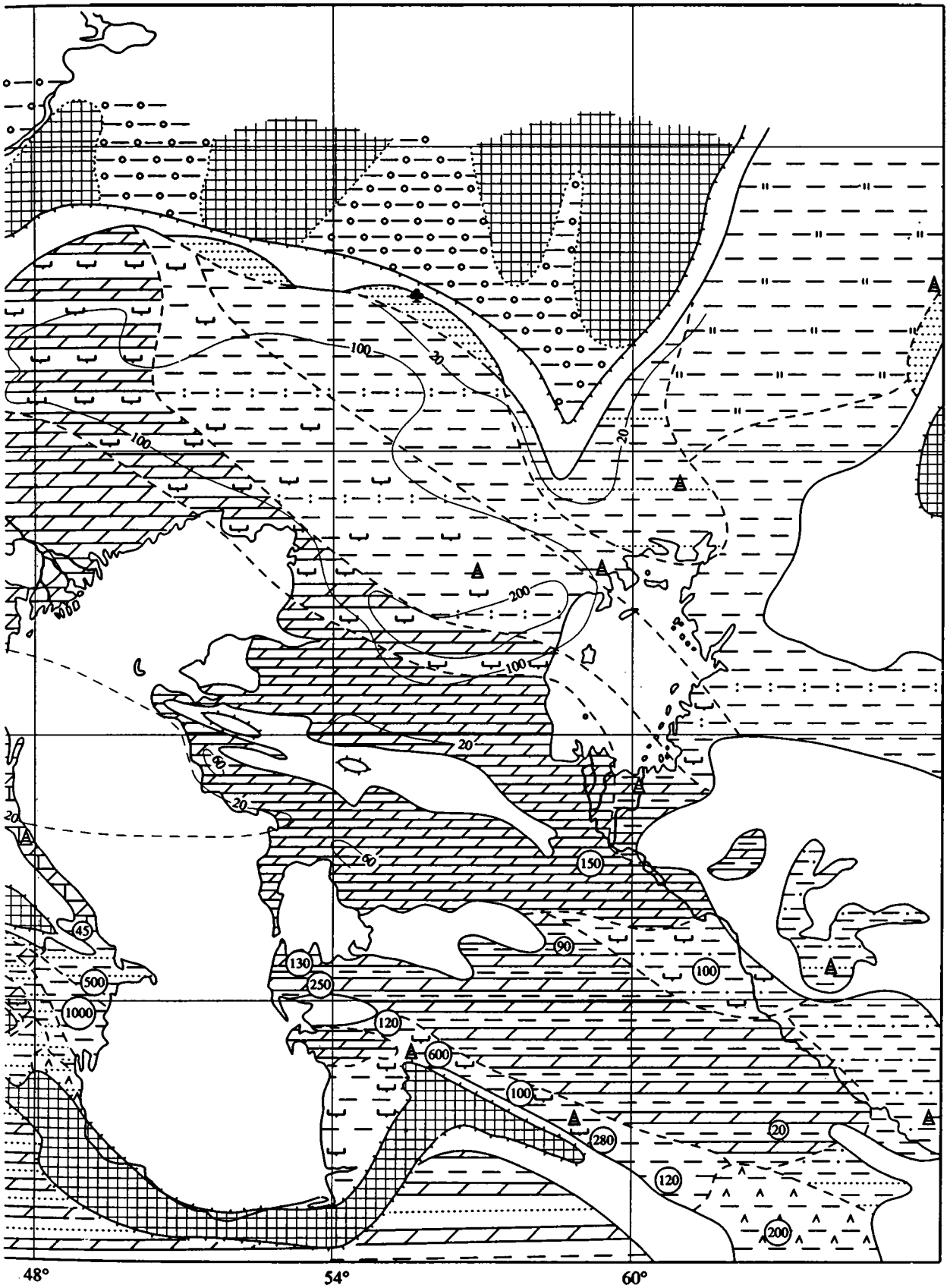
The hydrogen-sulfide contamination of the bottom waters in the early Oligocene was one of the most important hydrological events in the deep-sea zone of the Eastern Paratethys, because it completely annihilated all life in the bottom waters. The aerobic environment existed only in the marginal and shelf seas and in the upper parts of the slopes of the deep-sea basins.

These hydrological conditions favored the accumulation of higher concentrations of dissolved manganese in the hydrosulfuric zone, and its precipitation in the marginal parts of the deep-sea basins at the oxygen hydrochemical barrier. This process may be responsible for the great epoch of manganese ore formation in the early Oligocene (Nikopol', Chiatura, and Mangyshlak ore districts; the Varna district in Bulgaria). The impoverished composition of phytoplankton and the abundance of amorphous organic matter also indicate an oxygen deficiency in the deep-sea zone.

Organic life. The molluscs of the entire Pshekhian basin were represented by similar assemblages with a slight difference between the northern and southern (Transcaucasian) coasts of Eastern Paratethys. In relatively deep-sea conditions, the soft substrata were inhabited by assemblages of *Palliolum simile, Yoldiella chadumica,* and *Thyasira nysti,* which substituted the assemblage of *Propeamussium fallax.* The typical molluscs of the sandy substrata in the northern regions were *Pterolucina batalpaschinica, Astarte gracilis,* and *Cardites lukovichi.* In the southern part of the basin, the deep-water assemblages are unknown. On sandy substratum, the assemblages were dominated by *Glycymeris obovatus, Nemocardium tenuisulcatum, Pecten arcuatus,* etc.

During the second half of Pshekhian time, the basin regressed, and the sediments coarsened. Consequently,

Fig. 1. Late Eocene basins.
A – Lithofacies map of Beloglinian basin; B – paleogeographical scheme of basins in western Europe (See Legend to Fig. 5).


48°
54°
60°
100
20
200
150
90
45
130
250
500
1000
120
600
280

in the northern part of the basin, the assemblages of a coarser substratum developed extensively with dominations of *Arctica rotundata, Glycymeris lunulata,* and *Cardites kenderlensis.*

In the Pshekhian basin, no fewer than 175 species (Fig. 3) of bivalve molluscs were identified; the larger part of the fauna was composed of species inherited from the Eocene (68) and of migrants from the European basins (50). Most of the endemic species (36) appeared during the second half of Pshekhian time in the Transcaspian area. The salinity of the Pshekhian basin was almost normal (no less than 30‰), although it occasionally dropped to 15 - 20‰ at the northeastern margin.

The organo-walled phytoplankton of that stage belonged to no fewer than 100 taxons. However, the absolute majority of these taxons were found sporadically. On the whole, the share of phytoplankton, among all the palynomorphs, was much smaller than in Beloglinian time (5 - 10‰) The genus *Phthanoperidinium* became dominant due to its lower sensitivity to changes in salinity. A noticeable amount of the *Glaphyrocysta* species in the spectra is indicative of water stratification with higher temperatures in the surface layer.

In the Volga–Don region, in the second half of Pshekhian time, *Micrhystridium* was dominant in the phytoplankton assemblage, which implies the advance of the northern sea coast, the shallowing of the basin, and a more intense runoff of fresh waters.

Climate. The universal presence of such genera of molluscs as *Barbatia, Pteria, Crassatina, Nemocardium* s.s., and *Lutraria* in the entire Pshekhian basin is evidence of a rather high temperature, which was probably closer to that of the present-day subtropical basins than to the boreal basins. The paleobotanical data confirms this observation; the imprints of leaf flora from the northeastern coast of the Pshekhian basin (Ustyurt, North Aral) show a change from xerophilic, subtropical flora inherited from the Eocene to temperate mesophytic flora. But even at the time of the Ashcheairyk Formation accumulation, when Taxodiaceae and broad-leaved flora prevailed, a great number of thermophilic subtropical plants were growing in this area, e.g., Lauraceae, *Myrica, Comptonia, Magnolia,* etc. (Bronevoi and Zhilin, 1967; Rayushkina, 1983; Akhmet'ev and Zaporozhets, 1992). The composition of the spore-pollen assemblages of Ciscaucasia and Mangyshlak, supplied by the calm rivers of the plains, likewise indicates the presence of subtropical forests on the northern coast of the Pshekhian sea. The vegetation in the western and eastern parts of the northern coast differed slightly in their share of thermophilic plants. The climate of the eastern part, i.e., the Kazakhstan province, according to V.S. Kornilova (1966), was more arid, and desert plants grew there. On the whole, the findings of many researchers testify that the floral evidence shows a distinct botanical-geographical differentiation in the distribution of vegetative paleocoenoses since the beginning of the Oligocene.

The distribution of molluscs also indicates a rather definite climatic zonality in the early Oligocene and shows the differentiation of faunas by their relation to the temperature on the north and south coasts of the Eastern Paratethys. The Transcaucasian part of the basin alone was inhabited by the representatives of the genera now living in tropical and subtropical waters, such as *Atrina, Isognomon, Cubitostrea, Megaxinus s.s., Ctena, Anodontia,* and *Europicardium.* But astarts, cultellids, and sphenias are less numerous here, being more typical of the northern part of the basin.

The Solenovian Basin

Paleogeography and peculiarities of sedimentation. At the beginning of Solenovian time, the Paratethys lost its connection with the World Ocean, and a peculiar endemic fauna evolved. Though the basin closed, its size remained unchanged (Fig. 4), and locally, in the northern and eastern margins, the sediments were transgressive. In the inner parts of the basin (Ergeni, Chernye Zemli, and Ciscaucasia) the unconformable occurrence of the Solenovian horizon, with the erosion of underlying strata, was also frequently observed. There were large areas on the shelf, however, where calcareous deposits containing ostracods were absent (Mangyshlak, Northwestern Ciscaucasia) (Stolyarov and Ivleva, 1989). The link between the Eastern and Western Paratethys survived, and for the first time the Paratethys basin had a common fauna, as testified to by the finds of specific brackish-water molluscs in the middle part of the Tard clays of Hungary (Baldi, 1980, 1986), in the Bizusa beds of Transylvania (Rusu, 1983, 1988), and in the Socke beds of Slovenia.

Sharp changes in the hydrology of the basin caused the rearrangement of the water circulation and of the sedimentation environment. In large, relatively deep-sea areas, the dark, carbonate-free, silty-clayey sediments changed into light-colored, calcareous muds. The carbonates in these muds are unspecified, probably of biogenic origin, a product of the mass development and burial of a few species of plankton and, to a lesser extent, large ostracods (the "ostracod bed," or Polbinka beds). The basin became shallower, and the molluscs spread over its wide area, with the exception of, as earlier, the deep-sea off-shelf areas.

The desalinization of the sea in Solenovian time seems to reflect general changes in the salinity of the surface waters of the World Ocean (Muzylev *et al.,* 1992). This change in the salinity of the Paratethys is implied by the composition of the phytoplankton genera including *Gerdiocysta, Cyclonephelium, Distatodinium, Adnatosphaeridium,* and *Hystrichokolpoma.*

During the second half of Solenovian time, the connection of the Eastern and Western Paratethys with the open sea was restricted, as indicated by universal distri-

bution of several euryhaline marine immigrants, e.g., an impoverished assemblage of agglutinated foraminifers (Ter-Grigor'yants, 1964), the ostracods of marine transit species (Popov *et al.*, 1985), and molluscs. By that time, the conditions of sedimentation changed again, and the deposits were again represented by the typical "Maikop facies" over a larger area, i.e., the dark- and light-colored, carbonate-free clays with frequent pyrite nodules. These sediments undoubtably accumulated under specific bionomic conditions, which were usually unfavorable for molluscs. The environment, however, was not yet that of a typical hydrogen-sulfide contamination; the shelf deposits contain numerous remains of mud-eater burrows filled with sulfides or even red-brown iron oxides. The phytoplankton composition also indicates a weaker reduction environment and the basin-ocean connection at the time of the Ikiburul deposits. Only at the end of that period was the oxygen exchange again restricted during regression.

In Solenovian time, the basin became shallower and reduced locally, and on its periphery the shallow-water silty and sandy facies persistently accumulated (the Black Sea basin, Severnye Ergeni, Transcaucasia, and Kyzyl Kum). At that time, the uplifts in the west and central Caucasus had become the sources of clastic material, and the island chains with shallows inhabited by molluscs began to appear in certain localities (north Osetiya, northwest Caucasus).

Organic life. The molluscs of the early Solenovian (Polbinka) basin were represented exclusively by brackish-water and lagoon forms. Moreover, the genera and species of bivalve molluscs were endemic, whereas the gastropod species had wide geographical distribution, being identical to those of the lagoon and brackish-water assemblages of the Oligocene basins of Europe (Popov and Titova, 1982; Popov *et al.*, 1985).

More clearly than even marine fauna, the impoverished assemblages of brackish-water molluscs characterize, on the one hand, the unity of the mollusc fauna in the Paratethys and, on the other hand, the difference in its composition between the northern and southern coasts. In other words, most of the genera and some species of the fauna were identical, but the composition of molluscs of Transcaucasia was, by far, richer. For example, in the northern part of the basin, on the muddy substratum, the impoverished assemblages of *Ergenica cimlanica* and *Janschinella* dominated, and on the sandy sea bottom *Janschinella* and *Rzehakia usturtensis* dominated. In the Transcaucasian part of the basin, 20 species of bivalve molluscs were found with such typical forms as *Urbnisia lata* and *Janschinella*, and cardiids such as *Cerastoderma*, *Merklinicardium*, and *Korobkoviella* being prevalent. This autochthonous fauna evolved from only a few lagoon molluscs of the Pshekhian basin. The presence of transitional forms and the abundance of material allow us to trace the phyletic lineages for the greater part of the species and genera (Popov *et al.*, 1985). The salinity here probably reached no more than 10 - 12‰, with an altered ion composition; in the late Solenovian basin it was 12 - 14‰ with occasional increases to 20‰ in Transcaucasia. These salinity estimates are confirmed by the composition of the organo-walled phytoplankton: the cysts of the *Wetzeliella* and *Cyclonephelium–Adnatosphaeridium–Gerdiocysta* groups.

Climate. The endemic character of malacofauna supplies no data on the climate patterns of the Solenovian basin. The preserved differences in the diversity of assemblages between the northern and southern parts of the basin point only to the existence at that time of a sufficiently distinct climatic zonality. In most of the earlier papers, the Solenovian Stage was not treated separately, and the paleofloral data on this interval is therefore scarce. According to the spore-pollen data, the peak of the warm climate in the second half of Solenovian time is determined by the pollen of Fagaceae, including evergreen varieties, and by the extensive diversity of other thermophilic plants. In the coastal part of the southeastern framing of the basin, red-colored deposits appeared (the Saryin and Betpakdala formations) and the climate was more arid but still seasonal with a hot summer. Of the Turgai flora, the narrow-leaved sclerophyllous plants dominated, and some of them were still evergreen. V.V. Korallova (1984) wrote about the aridization of the climate at the time of the accumulation of the Molochanskaya and Serogozy formations in the Black Sea region. V.A. Borovoi and S.G. Zhilin (1967) noted a reduced number of thermophilic plants only in the upper part of the Chilikta Formation in the Aral area. The lower and middle parts of this formation correlate with the Solenovian Horizon through the discovery of *Ergenica* and *Janschinella*.

The Kalmykian Basin

Paleogeography and sedimentation. Many features of the general paleogeography and morphology of the late Oligocene (Kalmykian) basin of Eastern Paratethys resembled the setting in early Oligocene (Pshekhian) time (Fig. 5). Only the dynamics of sedimentation changed, and the depressions in the off-shelf areas began filling up with thick (over 1000 m) clayey sediments interbedded with sand and silt, which considerably compensated for the great amplitudes of subsidence. The most striking occurrence of "avalanche sedimentation" took place in the deepest part of the Eastern Paratethys in the vast depression of eastern Ciscaucasia and central Caspian, which has a typical uncompensated by sedimentation profile during the entire Paleogene (Stolyarov, 1991). The uncompensated relict depressions with thin sediments (100 m and less) survived only as narrow troughs to the north of the islands of the Greater Caucasus, whose paleogeographical interpretation is still uncertain.

Without doubt, significant paleogeographical rearrangement occurred in this area at the Oligocene–Miocene boundary. As a result, an island massif rose in the

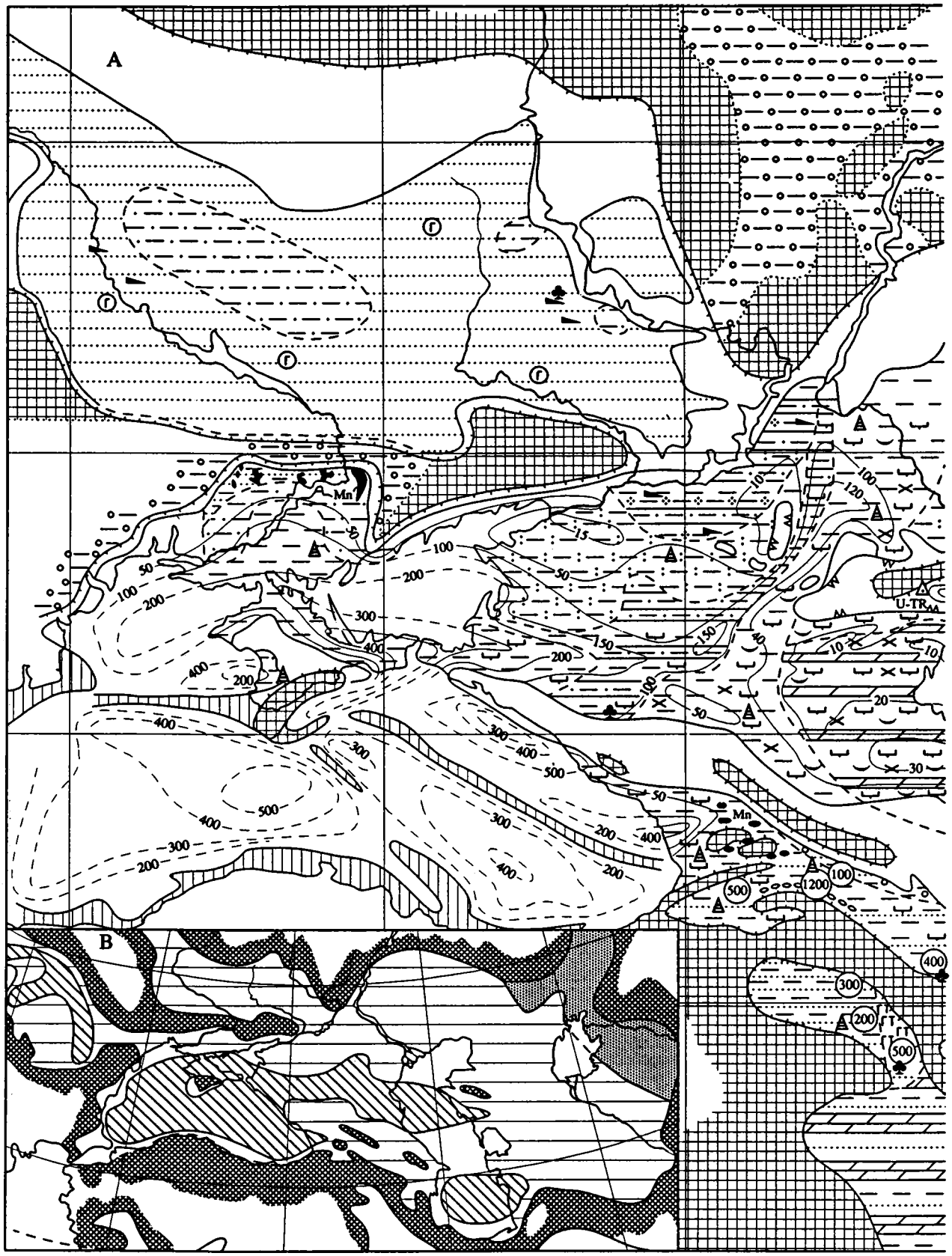
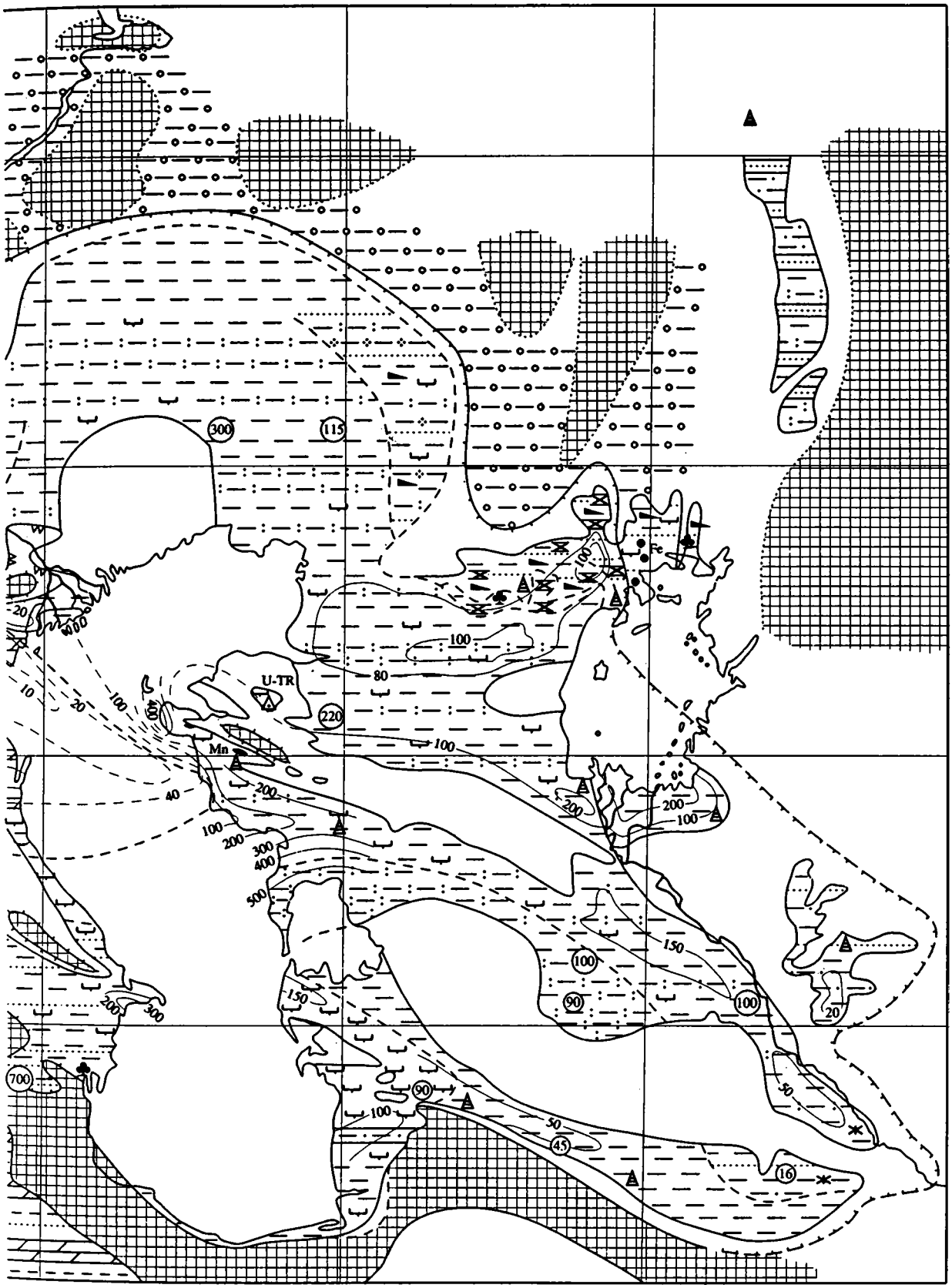


Fig. 2. Basins of the first half of the early Oligocene.

A – lithofacies map of the Pshekhian basin; B – paleogeographic scheme of the Paratethys and surrounding basins (See Legend to Fig. 5).


U-TR
Mn
Fe
300
115
220
700
16
45

eastern part of the Greater Caucasus, and supplied the basin with coarse clastic (sand-granule) material. Simultaneously, the previously uplifted eastern part of the Greater Caucasus subsided and was transformed into a deep-sea basin.

The northern barrier island boundary system was more distinctly pronounced than in the early Oligocene; it separated the deep-sea area from the intrashelf seas of the Dnieper–Donets, the Caspian depression, and the Aral regions. This island system can be traced from the Carpathians, across the Ukrainian crystalline massif, the Donets Basin, the Karpinskii Ridge, the Mangyshlak mountains, and the central Ustyurt uplift. This discontinuous system of low islands and shallows in the outer shelf area controlled the formation of metalliferous deposits of fish-bone detritus. These are, for the time being, unknown in other formations of the Earth and are unique in their origin and tremendous by scale (Stolyarov and Ivleva, 1989, 1991).

At the beginning of Kalmykian time, the connection between the Eastern Paratethys and the ocean broadened, and the basin was inhabited by marine fauna. The composition of the fauna and phytoplankton certainly points toward a renewal of the connection and a free exchange between the Eastern Paratethys and the Norwegian-Greenland basin, due to a great similarity in the taxonomic composition of dinoflagellates, molluscs, and ostracods (I.A. Nikolaeva, personal communications). The quest to locate the connecting strait was complicated by the absence of reliable upper Oligocene marine deposits in the territory of northern Ukraine, Belarus', and northern Poland. This connection may occur through the Cis-carpathian depression, assuming it was not a closed gulf at that time, or else through the Western Paratethys. The latter possibility seems unlikely due to the presence of warmer water fauna in the internal Carpathian basins and in the Rhein graben. Rusu (1977, Table 3) presents a cold-water assemblage of molluscs of which more than half are identical to those of the Eastern Paratethys for the beginning of Chattian time in Transylvania. The stagnant environment of the basin in the late Kalmykian is evidenced by the diversity of green algae and acritarchs, including *Cymatosphaera.*

Apparently, there was no direct paleogeographical connection between the Eastern Paratethys and the Tethys basin. Paleogeographic data shows that this is because there was land on the present-day territories of Turkey and Iran (Lüttig and Steffens, 1975). In the Western Paratethys, this connection is traced through Slovenia (Baldi, 1986), and the composition of the warm-water groups of fauna, which is recorded in Transcaucasia (the lower Uplistsikhe beds), suggests that they migrated from the east.

By the end of the Kalmykian, the hydrology and sedimentation were essentially different, i.e., the salinity deviated considerably from the normal marine salinity, and both sandy and sandy-silty deposits were more common. It is not clear whether the reduction in salinity was restricted to the northern part of the basin, or if it also involved the southern section. The unstable salinity and coarsening of sediments, however, do not imply the onset of regression, but in fact imply just the opposite. For example, in the Transcaspian part of the basin, the Baigubek deposits containing euryhaline marine fauna were more widely distributed than the older Karatomak deposits.

Organic life. The onset of marine sedimentation caused the extensive development of benthic fauna and the appearance of carbonate sediments in fairly deep areas. The most abundant of these were the benthic foraminifers of the genus *Virgulinella* (the "*Virgulinella* beds"). Later, however, the basin was contaminated by hydrogen sulfide, and the typical dark, carbonate-free, Maikop muds, rich in undecomposed organic matter with fish remains ("fish beds"), began to accumulate. The Maikop fish fauna became less diverse than that of the Pshekhian, the deep-sea and most of the warm-water, littoral fish disappeared, and the autochthonous species of the Maikop basin prevailed (Danil'chenko, 1980). Conditions extremely unfavorable for benthic fauna continued throughout the late Oligocene and the beginning of the Miocene when molluscs could inhabit only the margins of the basins. In the northern nearshore part of the basin, the soft, muddy substratum was inhabited by the assemblage whose composition was similar to that of the relatively deep-sea assemblage of the early Oligocene. The most widespread and typical genera were *Chlamys bifida, Palliolum, Thyasira,* and *Cuspidaria; Nuculana. Ch. bifida* was also the most widespread on the more silty substratum, whereas *Arctica rotundata, Pelecyora westendorpi, Glossus subtranversus, Pycnodonte gigantica callifera, Nemocardium cingulatum, Nuculana gracilis,* etc. were dominant on the sandy-silty and sandy substratum. The habitat depth of the shallow-water assemblages was a few dozen meters.

In the southern Transcaucasian part of the basin, the sand-gravel substratum was inhabited by *Glycymeris lunulata, Callista, Barbatia, Parvicardium,* etc. The salinity of the Kalmykian basin reached about 30‰, while in the marginal areas it fell to 15 - 20‰.

In the northern part of the late Kalmykian basin, the euryhaline species *Cerastoderma prigorovskii* and *Corbula* (*Lenticorbula*) *helmerseni* dominated in all kinds of substrata.

In the northeastern part of the basin, the salinity was unstable, varying from 25 to 5 - 10‰.

In all, only 90 species of bivalve molluscs have been recorded in the Kalmykian basin, apparently because of insufficient data for its southern part. The allochthonous species, similar to those of northwestern Europe, were dominant (58); most of the endemics species appeared in the second half of the Kalmykian (6 species and 9 subspecies).

The taxonomic diversity of dinocysts and prasinophytes of pre-Solenovian time was partially restored. The total number of species reached 120 - 150 in

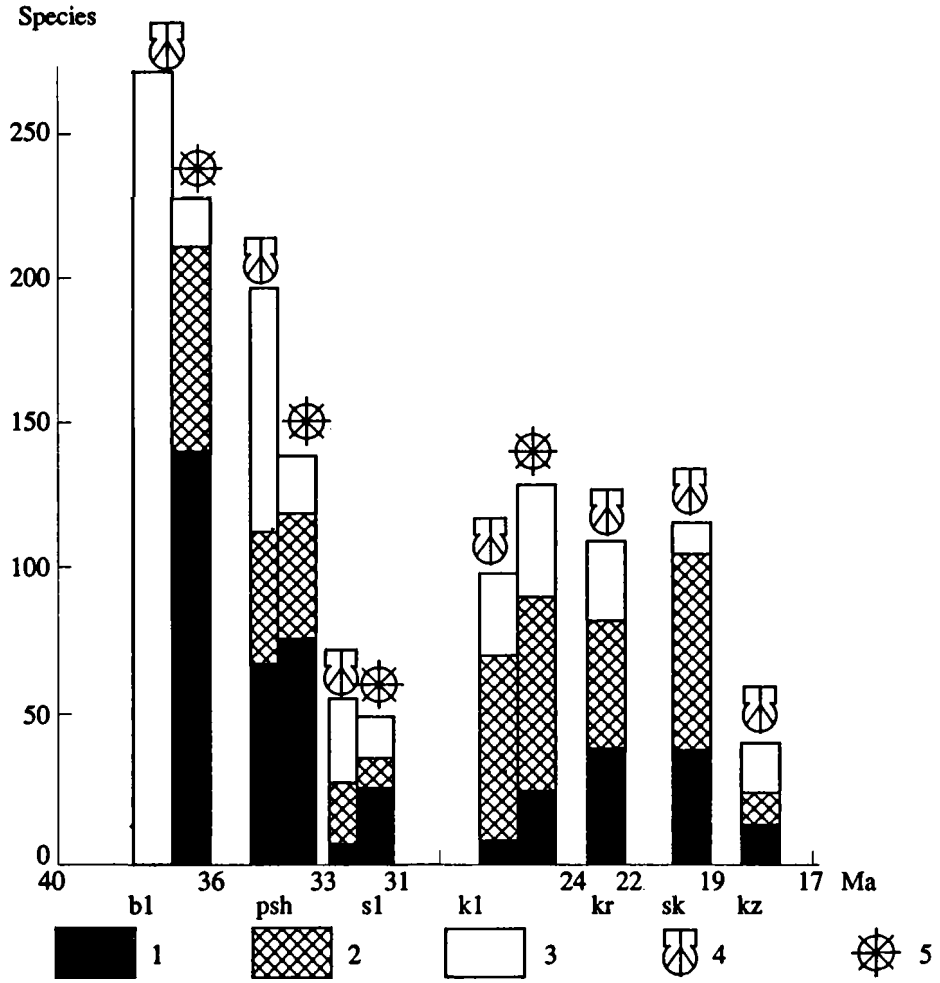


Fig. 3. A change in the taxonomic diversity of bivalve molluscs and dinoflagellates in the basins of the Eastern Paratethys.
1 – The share of species inherited from the previous basin; 2 – the share of allochthone species; 3 – the share of endemics appearing in the basin; 4 – bivalve molluscs; 5 – dinoflagellates.

the early Kalmykian and was then again reduced to 80 - 100 in the late Kalmykian. The variation in salinity in the late Kalmykian is revealed by the composition of the dinosysts, i.e., the larger part of the genera inhabiting the basins with normal salinity disappeared and was substituted by taxa that were more tolerant to salinity changes (*Deflandrea*, the species of the *Wetzeliella* group *Rhombodinium* and *Homotriblium*). In the vegetation of the coastal zone of the Paratethys, the mesophilic coniferous and broad-leaved forests were most common (the Turgai type, after A.N. Krishtofovich).

Climate. The almost similar diversity of faunas in the north Kalmykian and Pshekhian basins, and the genera of molluscs common to both show that there were only minor differences in the water temperature. And yet the water in the northern half of the Kalmykian basin was slightly cooler than in the early Oligocene, and representatives of subtropical genera were practically absent. The composition of the littoral fish fauna also indicates cooling (Danil'chenko, 1980). The spore-pollen data testify to the late Oligocene sea coasts being covered with Pinaceae, Taxodiaceae, and deciduous broad-leaved forests with rare subtropical plants: *Myrica, Magnolia,* etc. (Kozyar, 1959; Panova, 1981).

The spectrum of Amentiflorae was dominated by Betulaceae. This combination of vegetation was almost the same along the northern coast, evidencing a balanced, moderately warm, humid climate. In the Transcaucasian part of the basin, the presence of such genera as *Arca, Barbatia, Isognomon,* and *Arcopagia* implies much warmer waters and a subtropical environment. Floral data for the upper Oligocene of Ciscaucasia is scarce, but the macro- and microfloral data on Armenia and the northern slope of the Lesser Caucasus likewise indicates the presence of subtropical components in the floral assemblages.

The Karadzalgan Basin

Paleogeography and sedimentation. It was mentioned previously that the data on the deposits of this basin is extremely scanty and insufficient. If our corre-

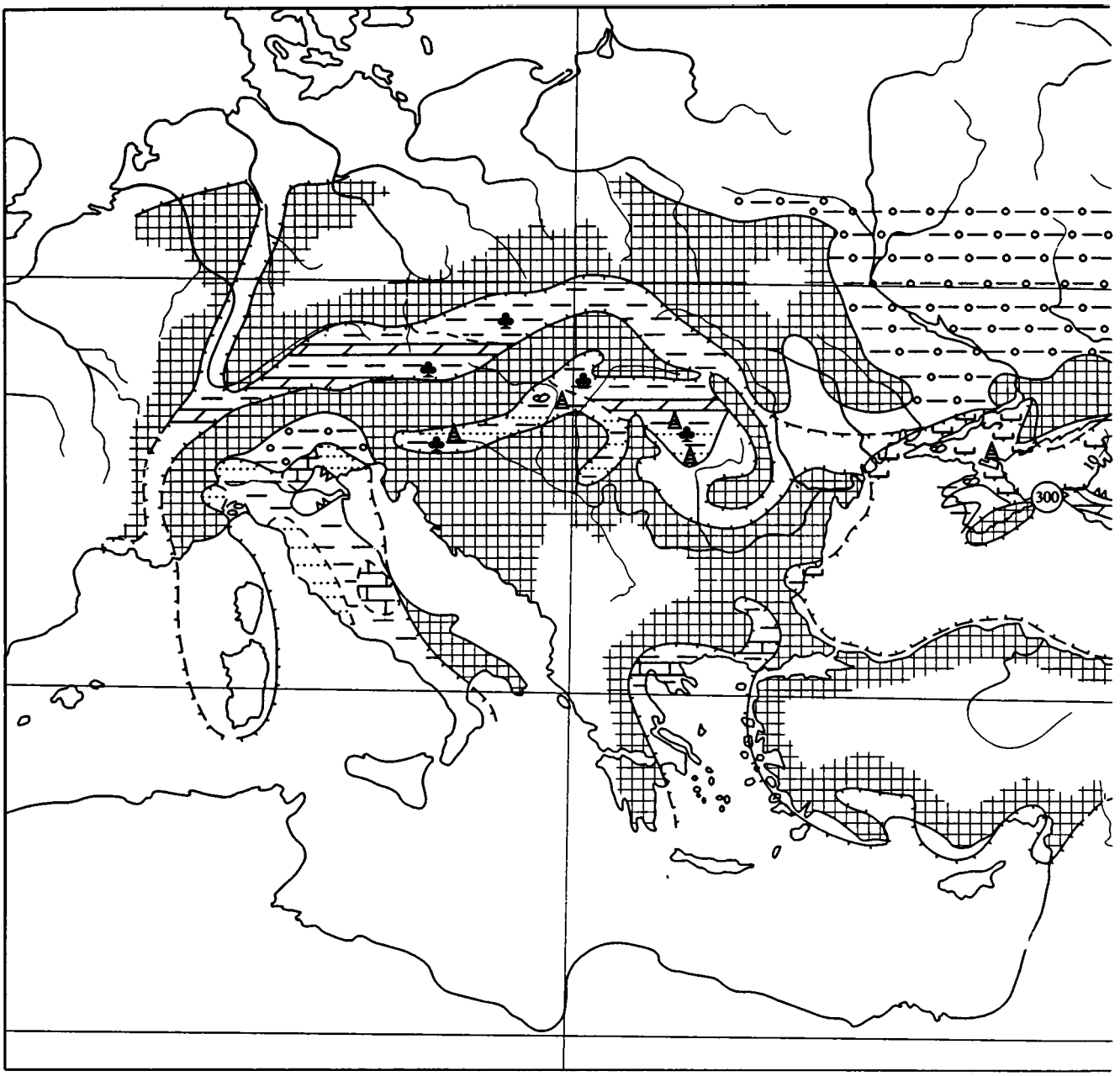
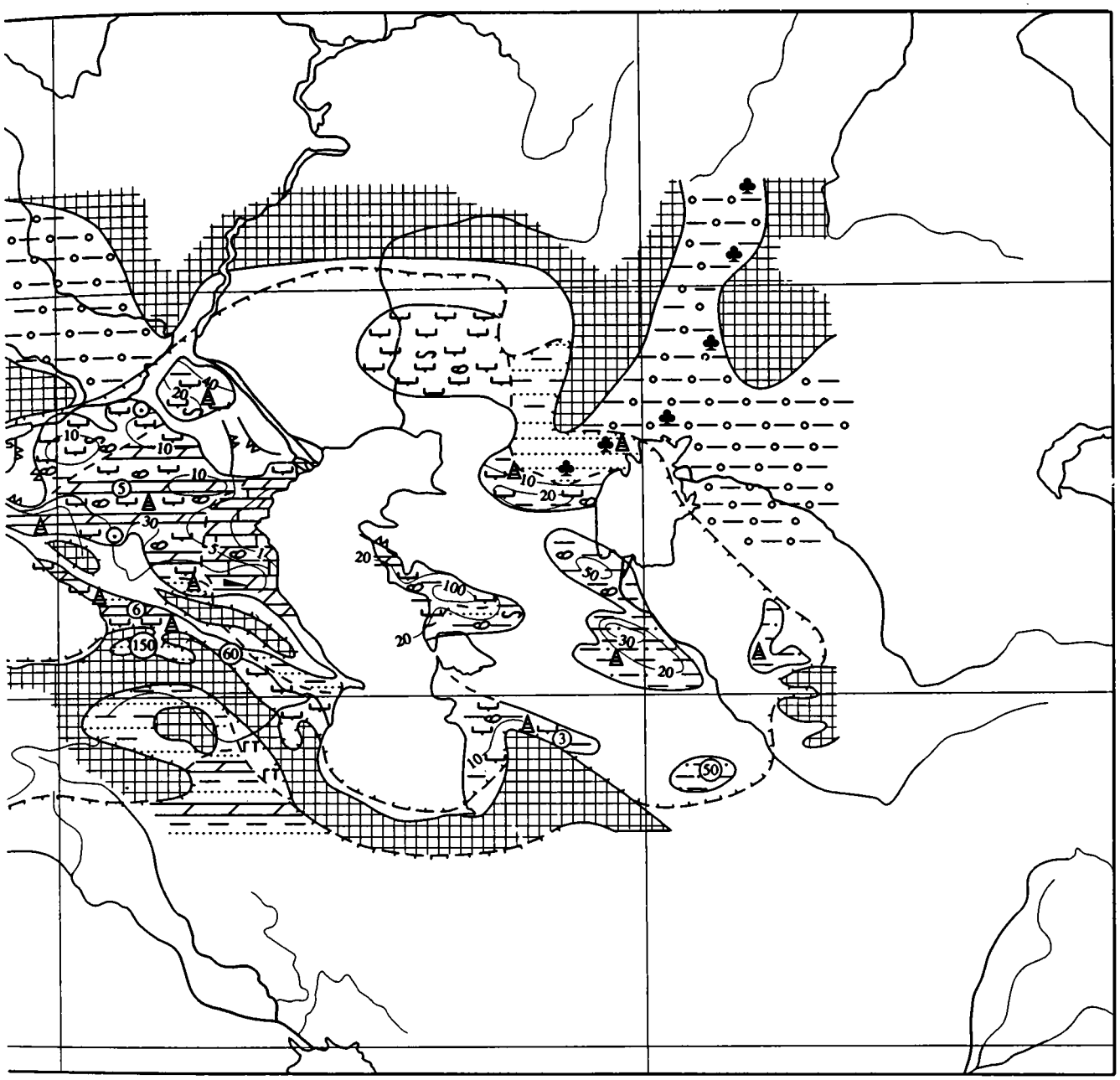


Fig. 4. The litholfacies map of the Solenovian basin (See Legend to Fig. 5 B).

lations of the Sivash beds of Ukraine, the Kintykcha beds of Ustyurt, the Kashkarata layers in Mangyshlak, the upper beds of the Sarbatyr Formation in Kyzyl Kum, the Kizylcha beds in Turkmenia, and the Gissarak beds in Tadzhikistan with the deep-water Karadzhalga Formation deposits of Ciscaucasia are correct, then the transgression of that period was strikingly apparent in the northwestern (Dnieper–Donets) and eastern margins of the basin, whereas no traces of it are found in the northern Aral area or in Ciscaucasia (Fig. 6). The general pattern of sedimentation has many features in common with that of the preceding stage of evolution. However the deepest (50 m), uncompensated by sedimentation part of the basin occurred in the eastern North Caucasus monocline; in the Terek-Caspian depression, farther to the north the thickness of the sediments reached 300 m.

In the marginal areas, quartz sands accumulated along with diverse, poorly preserved assemblages of molluscs. Together with many species inherited from the previous basin, numerous new warm-water immigrants applared, many of them similar to those of the Egerian fauna of central Europe. By that time, the connection with the Norwegian-Greenland basin was lost, possibly because of climatic barriers, and the main connection with the World Ocean was through the Western



Paratethys. In the deep-water zone of the sea, muds rich in undecomposed organic matter and, evidently, with hydrogen-sulfide contamination and abundant fish remains (the "fish" lithofacies) accumulated at that time.

Organic life. Molluscs. The silty muds of the northern margin of the basin (Ergeni), were again populated by polyhaline benthic forms: *Palliolum incomparabile, Lentipecten corneus, Nemocardium,* etc. On the sandy substratum in the Dnieper–Donets and Transcaspian depressions, *Callista lilacinoides, Pteria, Cubitostrea, Gibbolucina, Glycymeris,* and *Venus multilamella* were typical. The Transcaucasian part of the basin was inhabited by warm-water polyhaline genera: *Arca, Barbatia, Isognomon, Cardita s.s.,* etc. Salinity reached about 30‰, and probably dropped to 10 - 15‰ in the eastern part of the basin.

In all, about 100 species of bivalve molluscs were recorded in the basin, almost half of them (42) being species similar to those of the Oligocene basins, and a considerable part (39) were new migrants, mainly from the Western Paratethys.

The impoverished phytoplanktonic assemblages contained no more than 40 - 50 taxa. By this time, apparently as a result of cooling in the late Karadzhalgan, many Paleogene species became extinct, in particular the representatives of *Deflandrea, Thalassiphora,*

practically the whole *Wetzeliella* group, etc. On the land surrounding the Paratethys, large open spaces were occupied by herbaceous coenoses. The seasons were now more distinct, and the forest mesophilic Turgai flora became impoverished.

Climate. The presence of bivalve molluscs definitely indicates that the temperature of the warm Karadzhalgan basin was higher than that of the earlier Oligocene seas. At that time, a much richer fauna was introduced into the Eastern Paratethys and populated the region with subtropical genera new to the basin and with genera that earlier inhabited only its southern part, such as *Isognomon, Atrina, Cubitostrea, Megaxinus* (*Gibbolucina*), *Divalinga, Cardita s.s., Europicardium, Venus*, etc., Against the background of such mollusc diversity, the differences between the faunas of the northern and southern coasts seem less striking. However, this conclusion may be prompted by the paucity of data on Transcaucasian fauna. The Karadzhalgan sediments still contain pollen testifying to the development in the coastal zones of mesophilic coniferous and broad-leaved forests, in which the proportion of evergreen and sclerophyllous plants was small. The sediments of the Kashkarata Formation in Mangyshlak contain a large amount of pollen from thermophilic plants, including palms, laurels, myrtles, etc. In the Transcaucasian part of the basin, L.A. Panova (Voronina *et al.*, 1991) noted a considerable change in the pollen composition in our samples from the upper beds of the Uplistsikhe Formation. This is probably evidence of a change in paleogeography, which resulted in the appearance of typical Miocene taxa, rather than the direct consequence of a climatic change. In other words, the dissected topography became more apparent. For the Tadzhik depression at the time of accumulation of the Gissarak-Shurysai deposits, M.Z. Pulatova's data (1983) also implies a considerably rugged topography with well-defined vertical belts of vegetation, an increase over time of the proportion of herbaceous plants, and the xerophytization of the assemblages.

We may conclude that there was a drier climate at the Oligocene–Miocene boundary and at the beginning of the Miocene, when coniferous and broad-leaved forests were dominated by semi-open spaces that were covered with shrub-herbaceous coenoses. In the Ciscaucasian region, an increase of the dark-conifer species in the pollen assemblage is probably related to a prominent upraising of the Greater Caucasus at the time.

The Sakaraulian Basin

Paleogeography and sedimentation. The Sakaraulian basin was regressive, in particular on the northern and southern coasts, where at that time a hiatus, or continental deposits (Fig. 6), were formed. The marine environment survived only in the western Kopet Dag and in the individual depressions of the south Aral area. Regressive sediments were found in the Black Sea depression and in the Ciscaucasian area, where the Labinsk manganese ore deposit was formed in the delta beds on the northern slope of the west Caucasus island group. In the littoral areas, sediments changed from clayey to sandy (the Kartli depression, Western Georgia, Kopet Dat). For the most part, clayey sediments continued to accumulate in the deep-sea depressions.

Despite the regressive regime, the Sakaraulian basin probably had more extensive connections with the world ocean, and, without a doubt, the connection between this basin and the Western Paratethys was still wide enough to maintain an exchange of fauna between them and also with the Burdigalian basin of the Mediterranean further to the west. These connections are confirmed by both molluscs and ostracods, e.g., the sediments of the Ol'ginskaya Formation for the first time show the Mediterranean genus *Carinovalva* (Shneider, 1968). All peculiarities of Sakaraulian fauna, however, cannot be attributed to these connections with the ocean through the Western Paratethys, because among the Sakaraulian molluscs there are many obviously allochthonous marine forms, which are absent in European fauna. Such species are particularly numerous among cardiids, and some of their genera were more typical of the Indo-Pacific fauna than of the Mediterranean (e.g., *Plagiocardium* and *Fragum*). The fish fauna contain Indo-Pacific elements as well, e.g., the genus *Alepes* from the *Bathysiphon* Formation in the eastern Crimea (after Bannikov). These finds imply the introduction of migrants from the Indo-Persian area. This is supported by indications that, in Burdigalian time, this kind of connection was quite real: in the early Miocene there was a vast sea basin in Eastern Turkey and Iran (Lüttig and Steffens, 1975; Steininger *et al.*, 1985).

Organic life. Molluscs. In the northern part of the basin, the muddy bottom was inhabited by *Palliolum simile, Plagiocardium abundans, Nucula*, etc. Assemblages of coarser substrata were not recorded.

The southern (Transcaucasian) part of the basin, on the contrary, was inhabited exclusively by sandy-substratum molluscs, predominantly *Anadara sakaraulense, Acanthocardia kupradzei, Arctica, Glycymeris*, and *Fragum*. In all, about 100 taxa inhabited the Sakaraulian basin. Species inherited from the earlier Eastern Paratethys (38) and immigrants from the Western Paratethys (34) were prevalent. Moreover, the mollusc assemblages were considerably diversified by species, which were probably of Indo-Persian origin (17). All recorded assemblages were represented by polyhaline forms indicative of salinity that was close to normal.

The lower Miocene phytoplanktonic assemblage was impoverished and usually made up no more than 5% of the total palynomorph content. The species diversity was not high (hardly more than 50 species), however, the Miocene forms belonging to the genera *Tuberculodinium, Paleocystodinium*, and *Melitosphaeridium* were already conspicuous.

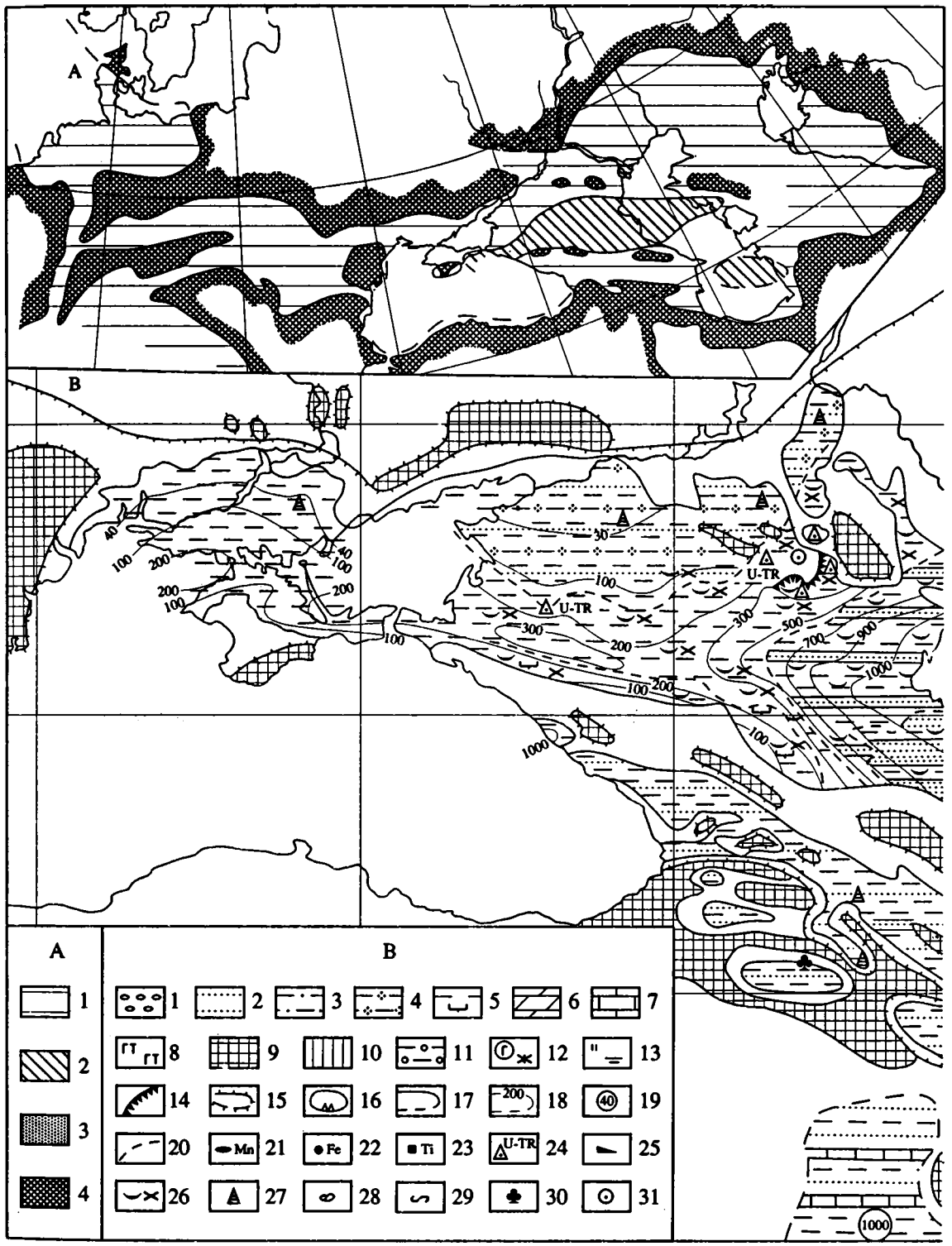


Fig. 5. Late Oligocene basins.

A – Paleogeographical scheme of the Paratethys and surrounding basins: 1 – shelf; 2 – deep-sea uncompensated basins; 3 – plain occasionally flooded by the sea; 4 – land. B – lithofacies map of Kalmykian basin: 1 – the granule-pebble deposits; 2 – sands and sandstones; 3 – silt and siltstones; 4 – sandy and silty clays; 5 – clays and calcareous clays; 6 – marls; 7 – limestones; 8 – igneous rocks, acidic and intermediate; 9 – eroded land; 10 – areas without sediments (Fig. 2 shows the early Oligocene map); 11 – accumulative lowland plain; 12 – rocks with glauconite and red rocks; 13 – siliceous and clayey rocks; 14 – benches of the sea bottom; 15 – shoreline, established and supposed; 16 – areas without sediments; 17 – contour line of sediment distribution today, established and supposed; 18 – isopachs, established and supposed; 19 – thickness of deposits in individual sections; 20 – boundaries of lithofacies. Mineral deposits. 21 – manganese; 22 – iron; 23 – titanium; 24 – uranium and rare metals; 25 – coal. Fossils: 26 – remains of fish: scales and bones; 27 – molluscs; 28 – ostracods; 29 – burrows of mud-eaters; 30 – flora; 31 – stratotype and parastratotype sections of regional stages.

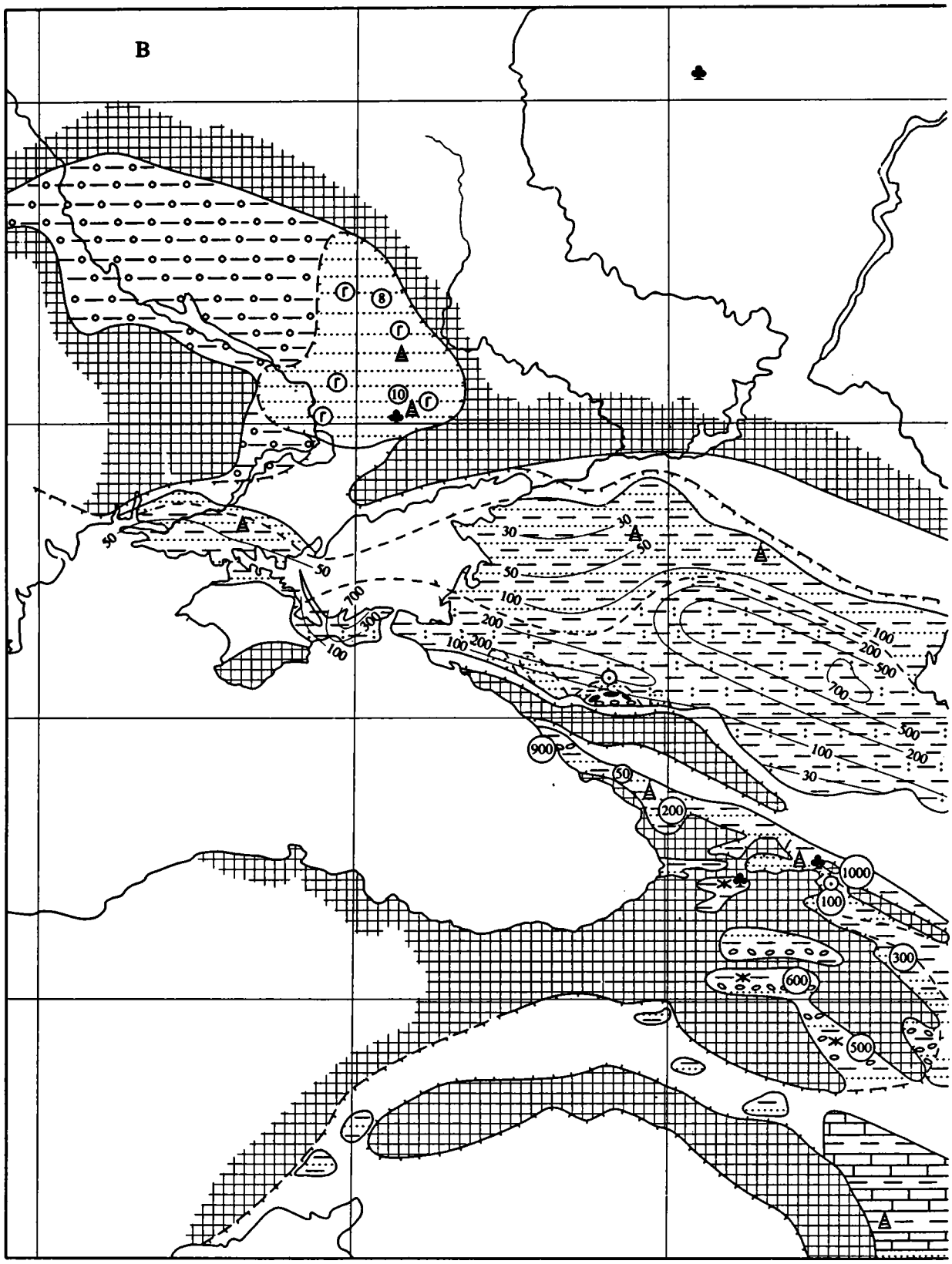
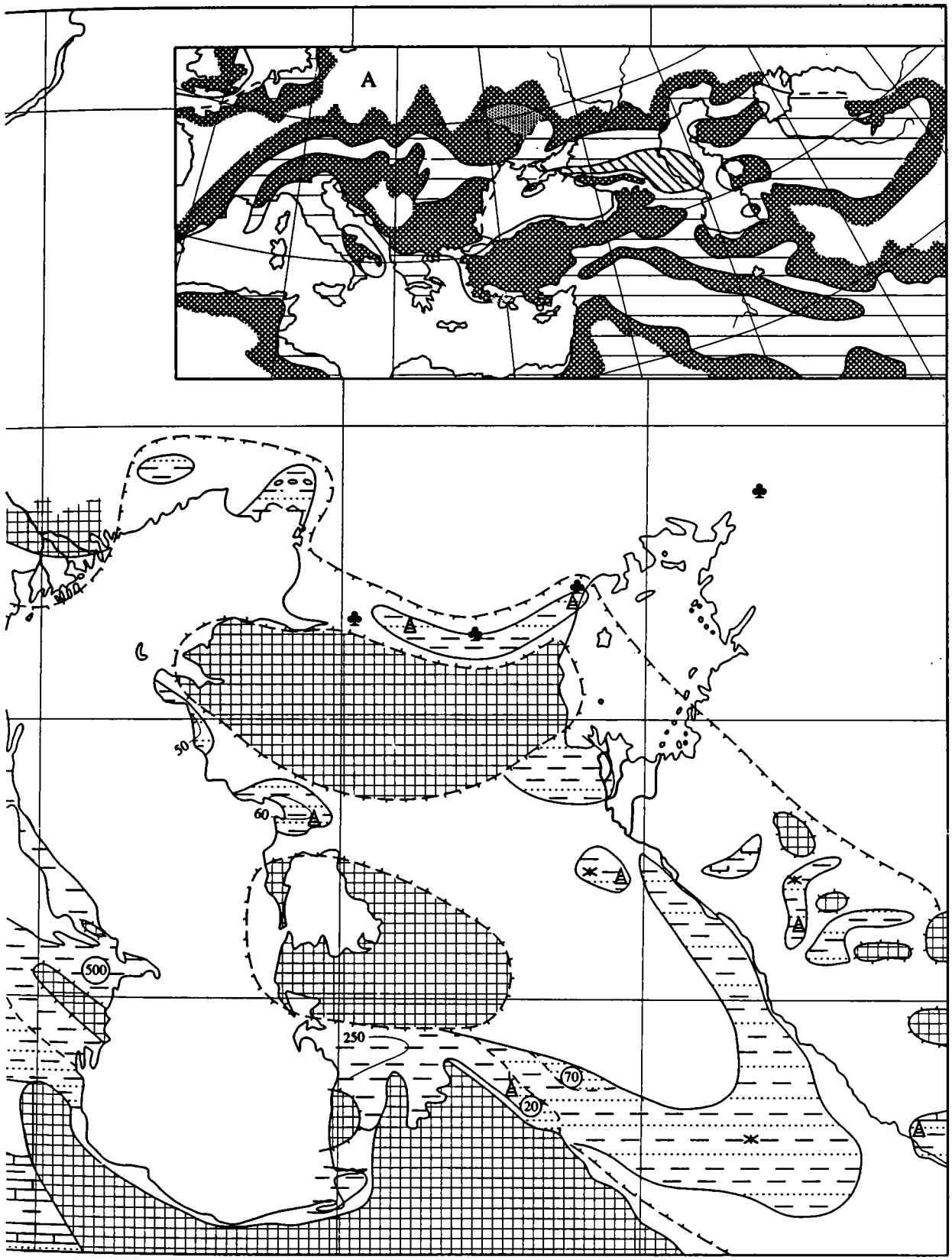


Fig. 6. Basins in the beginning of the early Miocene.

A – Paleogeographical scheme of the Paratethys and surrounding basins; B – lithofacies map of the Karadzhalgan-Sakaraulian basin.


A
500
50
60
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70
20

Climate. The Sakaraulian molluscs are the most thermophilic of all known faunas of the post-Eocene Eastern Paratethys. The subtropical forms of this fauna were for the first time accompanied by the species of the genera whose modern representatives now inhabit only tropical waters (*Fragum* and *Discors*). Moreover, the large size of the molluscs and the great diversity of assemblages also testifies to a very warm underwater environment. There is also palynological evidence of a sharp warming of the climate and an increase in vegetative diversity, i.e., our samples from the Sakaraulian part of the Gori–Uplistsikhe section of Panova (1981) show rich assemblages with diverse subtropical flora and numerous exotic forms that imply a seasonal, hot climate. On the northern seacoast, changes in the palynological assemblages were not so obvious. According to Korallova (1982), the phytogeographical unity of this area with the boreal floras of the North Sea coasts is still apparent in the dominance of mesophytic Arctotertiary coniferous and broad-leaved plants. To the east of that area, in addition to forests, savannah-type flora existed with xerophytic elements and numerous *Ephedra* (Zaklinskaya, 1958). Because of very dry climate, *Taxodiaceae* forests were substituted by pine forests (Stratigrafiya SSSR..., 1986). The coastal zone, which bordered the Paratethys in the east, was characterized by a seasonal, warm climate with dry summers. In this area, the semi-separated lagoons accumulated carbonate and clay–carbonate deposits of the Aral Formation s.l.

The Kotsakhurian Basin

Paleogeography. In the Kotsakhurian, as at the beginning of the Solenovian, the salinity of the water again decreased and the basin was inhabited exlusively by endemic, brackish-water fauna. The size and the configuration of the basin were only slightly different from those of the Sakaraulian basin.The sea was still regressing in the west and almost disappeared in the Black Sea depression, whereas in the northeastern margin (the Ustyurt part of the basin), the analogs of the Kotsakhurian deposits occur transgressively.

At that time, the Western Paratethys was the isolated regressive basin, divided into four water bodies (Ottnangien..., 1973). However, the presence of common, brackish-water forms in these basins and in the Eastern Paratethys implies episodic exchanges between them (Popov and Voronina, 1983).

Though the hydrology changed drastically, the sedimentation in the main basin was unaffected, and the lithology of its sediments is identical to that of the underlying deposits. Apparently, the basin was still contaminated with hydrogen-sulfide. In the littoral zones, sandy sediments were prevalent, as in the Sakaraulian time.

Organic life. The composition of molluscs of the northern margin was extremely impoverished. The species *Rzehakia dubiosa,* accompanied by rare cardiids, was clearly dominant. The presence of the specific brackish-water genus *Eoprosodocna,* and the very euryhaline lagoon genera of marine origin, such as *Cerastoderma, Corbula* (*Lenticorbula*), and *Siliqua,* in the Kotsakhurian basin is indicative of both the desalinization of the Kotsakhurian basin and of a change in the salt composition. The extinction of such euryhaline genera as *Mytilus* and *Parvicardium* at the beginning of Kotsakhurian time suggests an environment similar to that in the modern Caspian Sea, with a salinity that never exceeds 10 - 14‰. The organo-walled phytoplanktonic assemblage from the Kotsakhurian deposits remains unstudied.

Climate. The endemic character of the Kotsakhurian molluscs provides no data for climatic interpretation. Palynological data indicates that the climate was similar to that in the Sakaraulian time, though drier. The large proportion of herbaceous plants in the pollen spectra marks the initial stage of the development of steppe vegetation (Stratigrafiya SSSR..., 1986). East of the Paratethys, in the Aral and Turgai areas, the sclerophyllous forests including Fagaceae (chestnuts and oaks), ash trees, pistachio trees, *Pterocarya*, etc. The pollen assemblages from the Kushuk beds of Turgai contain various herbs and *Ephedra*. The climate at that time was probably similar to the modern climate in the eastern Mediterranean.

GEOLOGICAL EVOLUTION AND MINERAL RESOURCES

The scale of ore-formation in the Eastern Paratethys in the Oligocene–early Miocene was tremendous. This basin formed the largest Phanerozoic province of industrial manganese orebodies, which also included unique, sulfide–phosphorus uranium and rare-metal ore deposits represented by bedded accumulations of fish-bone detritus, and iron sulfides with rare earths, uranium, scandium, rhenium, nickel, cobalt, molybdenum, etc. Such industrial orebodies are recorded nowhere else in the world. Oolitic iron ores, gas, and oil also occur in this province.

At the present time, we suggest the following stratigraphic control of the ore mineralization (within regional stages):

(1) Pshekhian Stage: the manganese deposits of southern Ukraine (Nikopol'), Georgia (Chiatura), Mangyshlak, Bulgaria (Varna); the uranium and rare-metal occurrences in the Volga–Don area (Chernye Zemli) and the Transcaspian (Buzachi Peninsula); and the iron ores of the northern Aral area.

(2) Solenovian Stage: the iron ores of the northern Aral area.

(3) Kalmykian Stage: the uranium and rare-metal deposits of the Transcaspian (Mangyshlak) and Volga–Don area (Ergeni); the iron ore deposits of the northern Aral area, possibly in part regenerated.

(4) Sakaraulian Stage: the Labinsk manganese deposit in the Northern Caucasus and the uranium and rare-metal occurrences in Ciscaucasia.

Two stratigraphic units, the Pshekhian and Kalmykian stages, have the largest ore deposit concentrations and, thus, indicate unique ore-forming epochs, i.e., the early Oligocene manganese and the Late Oligocene uranium and rare-metal epochs.

Tectonic movements in the early Oligocene constitute a full cycle, beginning with large-amplitude intense subsidences and terminating with a relative leveling of the sea bottom. Because of the high mobility of the subsidence areas (deep-sea basins), the tectonic movements were differentiated. The immense outburst of manganese ore-formation coincided with a change in the direction of tectonic movements from subsidences to uplifts. As the bottom of the deep-sea basins rose, the sea invaded some areas in the Ukraine and Georgia, where manganese ores were often formed and deposited directly on the weathering crust of exposed crystalline rocks. The simultaneity of transgression and ore formation could be related to both the evacuation of water from the basins and the rise of deep, hydrogen-sulfide, manganese-bearing solutions into the shelf and littoral zones, where the manganese hydroxides precipitated at the oxidation (oxygen) barrier. This historical-geological model of manganese ore-formation provides positive evidence for the synchroneity of ore deposits in the regions of the Eastern Paratethys that are far removed from each other (Ukraine, Georgia, Mangyshlak, and Bulgaria) but are all confined to the base of the *Spiroplectammina carinata* Zone.

This stratigraphic position of the manganese ore deposits was confirmed by the study of organo-walled phytoplankton in the Nikopol' region and in Mangyshlak, which established that the accumulation of ores occurred in the interval of the *Wetzeliella symmetrica* Zone and in the transition to the *Wetzeliella gochtii* Zone (Table 2). The phytoplankton in this interval contained abundant *Micrhystridium*, which indicates a shallowing and a greater fresh water inflow, possibly with an additional supply of manganese-bearing components. The appearance of abundant *Deflandrea* in the phytoplanktonic assemblage and in the cysts of the *Wetzeliella* group that were tolerant to salinity may also imply frequent salinity fluctuations, which were mild enough for these taxa to survive. It should be noted that the formation of iron ores occurred in the littoral zone and in the lower reaches of the river valleys where estuaries were ingressed by the sea.

Thus, the early Oligocene epoch of manganese ore-formation is definitely related to active tectonic processes and to the subsidence of basins along deep-seated faults, which could have served as the channels for hydrothermal manganese-bearing solutions to enrich the hydrogen-sulfide waters with dissolved manganese.

The late Oligocene tectonic cycle, in its most important features, resembled the early Oligocene, but with a different uranium and rare-metal ore formation period. The distribution of these deposits are more restricted, i.e., they are mostly confined to the northern periphery of the deep-sea basin of eastern Ciscaucasia and the middle Caspian (South Mangyshlak, Volga–Don) and to the system of island uplifts and shallows on the outer shelf. The formation of these unique deposits is even more mysterious because it involves biological phenomena such as the accumulation of living fish, their mass death, and the later reworking and concentration of their biomass into layers, or lenticular bodies of fish-bone detritus up to 20 - 30 km long and several meters thick. The bone mass is cemented by clayey-sulfide material, which, in addition to pyrite (up to 20 - 25%) contains rhenium, nickel, cobalt, molybdenum, and scandium. The bone phosphate contains concentrations of rare-earth elements, uranium, and scandium. The absence of geological analogs to these ore formations naturally impedes the search for a solution to the genetic problems. Without doubt, their formation, as well as that of the manganese deposits, is related to (1) paleotectonic, paleogeographic, and historical geological peculiarities of the deep-sea area of the Eastern Paratethys; (2) hydrogen-sulfide contamination of the water, which could be the cause of both the mass mortality of the fish and their preservation in a fossilized state containing concentrations of different ore elements.

In the marginal parts of the basin, the oolitic iron-ore deposits (northern Aral area) were formed from a different, "crustal" source of metal (Formozova, 1959). Moreover, the process of ore formation was active here throughout the Oligocene, whereas in the periphery of the deep-sea area, this process occurred during a relatively short period of time.

CONCLUSION

(1) We have studied sections of the upper Eocene and Oligocene–lower Miocene (Maikop) deposits of the Black Sea depression, Ciscaucasia, the Volga–Don region, Transcaucasia, the Transcaspian, Ustyurt, and the Aral regions with reference to their lithology and their fossil organic remains (molluscs, dinocysts, spores, pollen, and leaf flora).

(2) A unified stratigraphic scale of the Oligocene–lower Miocene in the Eastern Paratethys was worked out, which takes into account all new data on geological events.

(3) Litholofacies maps and paleogeographic schemes were compiled for the late Eocene (Beloglinian), the early Oligocene (Pshekhian and Solenovian), the late Oligocene (Kalmykian), and the early Miocene (Sakaraulian) basins.

(4) With greater attention to detail than in previous works, the environment of the basins were reconstructed showing the hydrology and the transgressive-regressive cycles of eustatic oscillations of the sea

level, the temperature regime, and the climate on the surrounding land.

(5) The evolution of biota of the Eastern Paratethys in the Oligocene was controlled by connections, as in the Neogene, with the World Ocean. However, the basin was, apparently, never completely closed, even during the Solenovian when its water exchange with the ocean was greatly restricted.

(6) The Maikop basin differed radically from later basins by the sharp tectonic and morphological differentiation of its separate parts and by the formation of uncompensated depressions in the early Oligocene and, to a lesser extent, in the late Oligocene. The deep-sea basins, in combination with intense river run-off, produced a distinct water stratification, recurrent stagnation, and the accumulation of tremendous volumes of undecomposed organic matter.

(7) These peculiarities in the Maikop basin caused the formation of unique deposits of useful minerals: manganese ores in the early Oligocene, uranium and rare-metal accumulations in the late Oligocene, and oolitic iron ores mostly in the mid-Oligocene. All deposits are distinctly, paleogeographically confined to certain elements of topography and facies.

(8) The phases of a warmer climate and the enhancement of its continental characteristics with clearly defined seasons coincided with the periods of restricted connection between the Paratethys and the World Ocean. During the periods of stable connections with the Tethys, the climate became warmer and more humid.

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Paleobotanic Characteristics and Radiometric Dating of Neogene Deposits in Southern Primor'e

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Abstract – The spore-and-pollen spectra from the stratotype sections of the Tavrichanka and Khanka horizons (Rechnoi Pen., Rettikhovka) were determined to be of Paleogene age, which does not correspond to the formerly adopted Miocene age of the units. According to geological data, palynological and diatom analyses, and fission-track datings, the Miocene in this region should be represented by tephroid strata. These strata are lithologically diverse (gravel, sand, and siltstone with lignite and brown-coal lenses) and include abundant pyroclastics. The dates obtained encompass an interval of 10.8 - 25.3 Ma ago. Sediments corresponding to the Miocene climatic optimum have not been found, possibly due to the incomplete sections studied.

Miocene continental deposits, filling the tectonoerosional basins of Southwestern Primor'e, were subdivided into three units, corresponding to the following regional horizons (from the base upward): the Tavrichanka (first half of the lower Miocene), Khanka (second half of the lower Miocene to the middle Miocene), and Ust'-Suifun (upper Miocene). This subdivision was stated in the Resolutions of the Interdepartmental Stratigraphic Conference held in 1978 in Vladivostok (Resheniya..., 1982).

The stratotype of the Tavrichanka Horizon is the lowest (sandstone and shale) member of the Ust'-Davydovka Formation exposed in the bluffs of the Rechnoi Peninsula in the Vladivostok vicinity. The coal measures exposed in the Rettikhovka brown-coal quarry are acknowledged as the stratotype of the Khanka Horizon. The lignite unit of the Ust'-Davydovka formation was recommended for placement at this horizon level (Resheniya..., 1982). The Ust'-Suifun Horizon has an areal stratotype situated in the lower reaches of the Razdol'naya River.

However, such stratigraphic schematics are not accepted by all investigators. In particular, M.A. Akhmet'ev, G.M. Brattseva (1973), and N.S. Gromova (1984), considering flora-bearing beds as the main object of paleobotanic study, argue that the Rettikhovka coal measures are of an Oligocene age. T.N. Bolotnikova (1988) correlates them with the upper Eocene to lower Oligocene deposits of the Daubikhe (Arsen'ev) basin. The Paleogene age of the Tavrichanka and Khanka horizons in the Rechnoi Peninsula (Ust'-Davydovka Formation) was suggested by B.M. Shtempel' (1926), R.S. Klimova (1971), and M.A. Akhmet'ev (1974). In one of the latest works on the Neogene stratigraphy of Primor'e (Kundyshev and Verkhovskaya, 1990), a definite opinion of the Eocene postoptimum age of the deposits in the section of the Rechnoi Peninsula was given. Without analyzing these publications in detail, we would like to note that sufficiently representative palynological materials, obtained from sections of the Rettikhovka and the Ust'-Davydovka Formation in the Rechnoi Peninsula, despite some regional differences, unambiguously indicate that these deposits are of Paleogene age. This is supported by the large role of various representatives of the Fagaceae family in the spectra, including the presence of the species of the *Quercus* genus, typical of the Paleogene: *Q. graciliformis* Boitz., *Q. gracilis* Boitz., *Q. conferta* Boitz., and *Q. forestdalensis* Tr., *Quercites sparsus* (Mart.) Samoil. (the content in the spectra from solitary grains to 3.6%). The spectra also include the pollen of *Fothergilla, Corylopsis, Hamamelis, Sycopsis, Disanthus, Laurus,* and *Platanus* genera (up to 4.9% of individual representatives of the group), as well as the pollen identified according to the artificial classification and characteristic of Paleogene deposits: *Triatriopollenites aroboratus* Pfl., *T. confusus* Zakl., *T. plicoides* Zakl., *Tricolpopollenites liblarensis* (Th. et Pfl.) subsp. *fallax* (R. Pot.) Th. et Pfl., *Tricolporopollenites cingulum* (R. Pot.) Th. et Pfl. subsp. *fusus* (R. Pot.) Th. et Pfl., *T. pseudocingulum* (R. Pot.) Th. et Pfl., and *Rhoipites* sp. (mainly solitary grains).

Therefore, there are grave reasons to doubt that the Tavrichanka and Khanka sediments belong to the Miocene. The stratigraphic position of the Rettikhovka and Rechnoi Peninsula sections is not the only problem. The widely assumed opinion that these sections are the same age as the coal measures and lignite-bearing strata in other basins of the western Primor'e complicates the situation. This opinion was shared by many researchers, including some writers of this paper (Pavlyutkin *et al.*, 1984). However, at present this idea clearly contradicts new data obtained through palynological and diatom analyses, and fission-track datings.

In addition to the Rettikhovka and Rechnoi Peninsula deposits, we studied a number of sections in the

Pushkinskoe, Vozdvizhenka, and Chikheza (Pavlovka) basins (Fig. 1). The borehole cores and materials from the quarries were used, applying lithological correlation, the study of diatoms, pollen, and spores, and the fission-track dating of ash interbeds. The results obtained indicate that the age of these sections is clearly different from that of the stratotypes of the Tavrichanka and Khanka horizons, and that they probably correspond to the Miocene stage in the geological history of Primor'e. Their sediments were identified as a stratigraphic unit of the so-called tephroid strata (Pavlyutkin *et al.*, 1988). Characteristics of these strata are: lithological diversity (from pelites to coarse psephites), an abundance of pyroclastics, an unusual discoloration from blue-green fresh rocks in cores to yellowish-white after their long exposition, a presence of thin interbeds of brown coal and lignitized wood remains, and, finally, a rapid lateral change of rock lithology. These features made it possible to correlate the sections during the field investigations. The correlation, with regard to laboratory data, is shown in Fig. 2.

The study of numerous individual sections shows that the tephroid strata in the basins of southwestern Primor'e overlie formations of various ages, with a stratigraphic hiatus and an erosional unconformity (from Cretaceous siltstones and sandstones to coal measures and lignite-bearing strata of the Ust'-Davydovka Formation). In turn, these strata are overlain by either the Pliocene–Quaternary deposits, mainly alluvial, or the upper Miocene gravel.

The tephroid strata consists of weakly cemented tuffstone and tuffaceous siltstone, and of vitroclastic tuff with interbeds of volcanic sand. Inequigranular polymictic sand and gravel are also present in the sections. Some intervals are enriched with lignitized organic matter, and valuable spore-and-pollen spectra

Scale 1 : 2500000

Fig. 1. Schematic map of the studied sections.

were recovered from them. The brown-coal beds are not numerous, and their role in the structure of the strata is insignificant. According to the refraction index, the volcanic glass corresponds to rhyodacite (1.495 - 1.500). According to the lithological analysis, its content in some interbeds reaches 80%. Assuming an insignificant transportation by streams from eruption centers and the negligible probability of redeposition in the course of the subsequent cycles of erosion and denudation, which was confirmed by the fresh appearance and poor roundness of the vitroclasts as well as by the lithological analysis of overlying sediments, the glass was used for fission-track dating.

The monofraction (>0.1 mm in size) was used for the fission-track dating of volcanic glasses. The tracks were chemically revealed by processing with concentrated HF at 21°C during 35 s, which resulted in the formation of acute-angled etching pits along the tracks. In dating all the samples, the population method (Gleadow, 1981) was used. The density of the induced tracks was determined by subtracting the density of spontaneous tracks from the total track density. This method of glass dating was chosen because preliminary annealing of spontaneous tracks could have resulted in a change in the chemical parameters revealing induced tracks after annealing the samples and in the cracking of glass fragments at high temperatures (Ganzei, 1981).

Smaller average sizes of spontaneous tracks in comparison with induced ones were noted in all the analyzed samples. This indicates a decrease in the density of spontaneous tracks as a result of their natural healing during the geological life of minerals. The original densities of spontaneous tracks were determined by a widely used method of correction "according to sizes" (Storzer and Wagner, 1970). All the age determinations have been made using the spontaneous decay constant of uranium $\lambda_f = 7.03 \times 10^{-17}$ years (the table).

It should be mentioned at once that the exact geochronological position of the lower boundary of tephroid strata is unknown, and, besides, it is probably not the same in individual sections. The relation of the strata to the underlying late Eocene coal measures was observed in a ditch at the Pavlovka-2 brown-coal quarry, outcrop (oc) 9035. Palynospectra of the same type were recovered from the lower tephroid strata. In these spectra, angiosperm pollen prevails over gymnosperm pollen, the amount of spores being insignificant. Species of the Betulaceae family distinctly dominate in the angiosperm group: *Betula* (4 - 19.3%), *Alnus* (11.6 - 25.3%), *Corylus* (0.9 - 3.0%) and *Carpinus* (2.3 - 3.9%), the presence of Juglandaceae pollen being rather noticeable (*Juglans* up to 4.2% in some samples). Characteristics of the Fagaceae pollen are: a very high content of the chestnut pollen (5.4 - 30%), an extremely insignificant presence of beech pollen, and solitary grains of oak pollen. The gymnosperm group is characterized by a predominance of the Pinaceae family with an uneven content of the Taxodiaceaea pollen (1.3 - 15%). *Pinus* subgen. *Haploxylon* clearly prevails among Pinaceae (up to 24.4%), while the dark conifer-

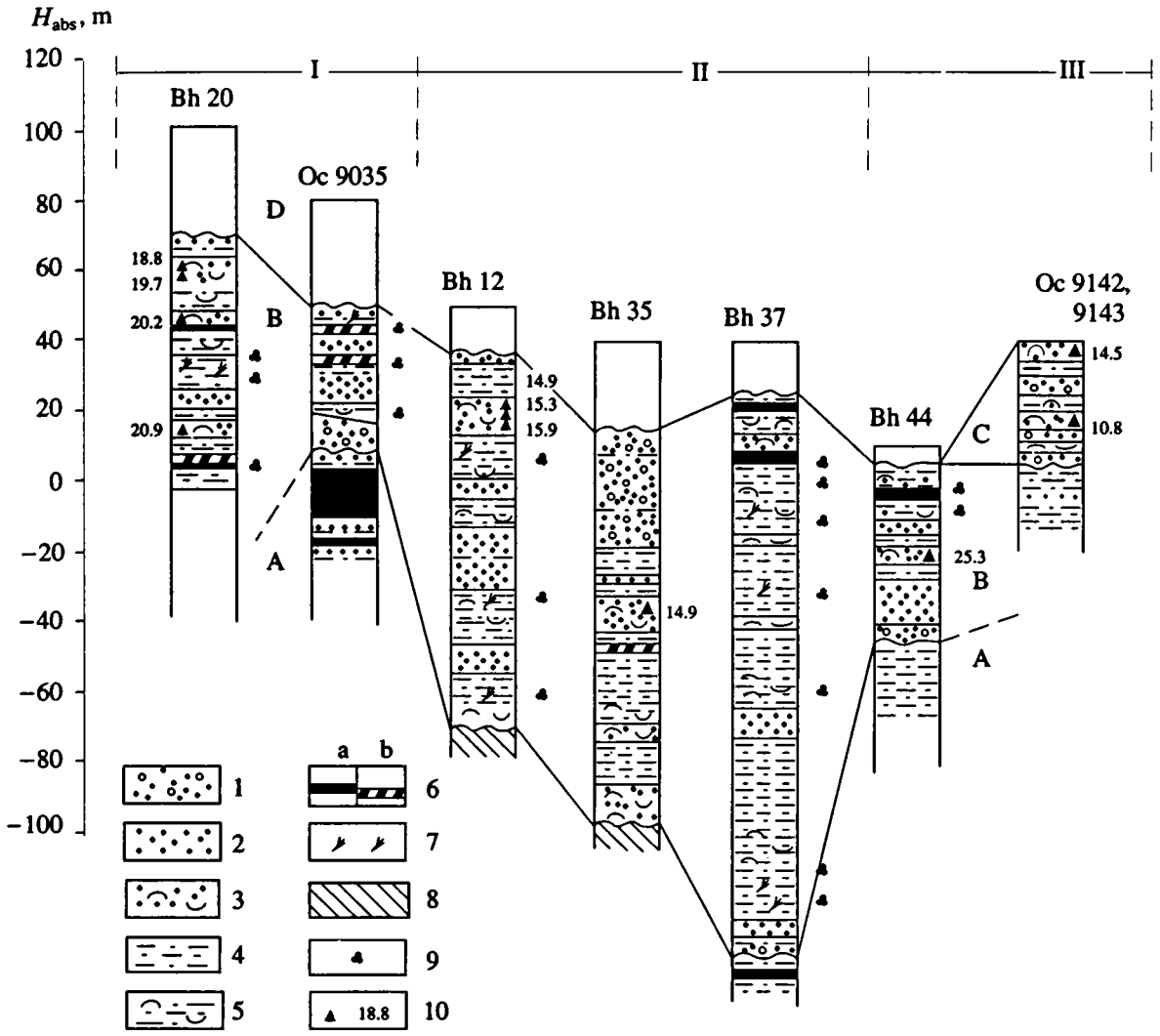

Fig. 2. Schematic correlation of the studied sections.

1 – Gravel; 2 – sand; 3 – volcanic sand; 4 – siltstone; 5 – tuffaceous siltstone; 6 – coal (a), lignite (b); 7 – plant remains; 8 – pre-Cenozoic rocks; 9 – samples for palynological analysis; 10 – fission-track age, Ma.

A – Pre-Neogene formations; B – tephroid strata; C – rudaceous deposits of the upper Miocene; D – Pliocene-Quaternary formations.

Depressions: I – Pavlovka (Chikheza); II – Vozdvizhenka; III – Pushkinskoe.

ous flora components are modestly represented. The pollen of the plants, indicators of Paleogene assemblages recovered from the underlying coal measures, is completely absent in the spectra.

Only a very fine (muddy) fraction of volcanic glass, useless for fission-track dating, was found in this section.

The younger tephroid strata were penetrated by borehole (bh) 44. The content of the Betulaceae pollen in their palynospectra decreases in comparison to the above-mentioned spectra. The beech pollen content increases (10.2%), while the chestnut pollen is completely absent. As in the spectra from outcrop 9035, no pollen from plants typical of the Paleogene was found.

Seven samples of diatom flora from the core of borehole 44 were studied. Well preserved diatoms, belonging to 22 species and 10 genera, were found in the samples. *Aulacoseira* (7 species), *Actinella* (3 species), and *Eunotia* (2 species) genera are most representative. Taxonomically the diatom flora is rather poor, which is common in early stages of the Miocene evolution of diatoms (Jouse, 1952; Moiseeva, 1967, 1974, 1979; Kuklina and Moiseeva, 1980). The main part of the diatom assemblage consists of *Aulacoseira praegranulata* (Jousé) Moiss. var. *praegranulata* + var. *curvata* (60%), *A. jouseana* Moiss. (8%), *A. distans* var. *ovata* Iwahashi (20%), *Actinella brasiliensis* Grun. (12%), *Desmogonium guianense* Ehr. f. *antiqua* Lupik. (4%), *Eunotia clevei* Grun. (4%), and *Actinella punctata* Levis (8%). In addition, *Aulacoseira praedistans* Jouse var. *praedistans*, *A. areolata* Moiss., *Melosira undulata* (Ehr.) Kuetz., *Paralia arenaria* (Moore) Moiss., *Biddulphia levis* Ehr., *Tetracyclus elliptici* var. *latissimus* Hust., *Fragilaria araulensis* Herib.,

Radiometric (fission-track) age of volcanic glasses from the tephroid strata

Sample no.	Density of spontaneous tracks $P_s \times 10^4$, tr/cm^2	Number of spontaneous tracks	Density of induced tracks $P_i \times 10^5$, tr/cm^2	Number of induced tracks	Neutron flow $\Phi \times 10^{15}$, tr/cm^2	Age (T), Ma	Preservation of tracks under natural conditions		Error in the age calculation, %	Corrected age (T), Ma
							size	density		
12-25	2.4093	461	1.9849	629	1.59	11.0	0.891	0.74	6.13	14.9 ± 0.9
12-30	2.4902	778	1.9053	667	1.63	13.4	0.974	0.88	5.28	15.3 ± 0.8
12-35	1.9728	697	1.6415	983	1.68	12.2	0.908	0.76	4.96	15.9 ± 0.8
91-43	1.6116	351	1.6547	1050	1.83	10.7	0.890	0.74	6.17	14.5 ± 0.9
35-70	1.4622	339	1.2982	780	1.89	12.8	0.960	0.86	6.50	14.9 ± 1.0
20-41	1.3766	498	1.1141	685	1.72	12.8	0.860	0.68	5.91	18.8 ± 1.1
20-43	1.3296	544	1.0377	658	1.74	13.4	0.869	0.68	5.78	19.7 ± 1.1
20-53	2.1471	674	1.3103	1098	1.77	17.4	0.960	0.86	3.85	20.2 ± 0.8
20-81	2.6831	581	1.8026	835	1.79	17.8	0.958	0.85	5.40	20.9 ± 1.1
44-36	2.1471	400	1.5008	667	2.03	17.4	0.867	0.69	6.32	25.3 ± 1.6

F. spinosa Skv., *Actinella pliocenica* var. *tenuissima, Desmogonium rabenhorstianum, Eunotia praerupta* Ehr., *Navicula americana* Ehr., and *Pinnularia borealis* Ehr. have been found.

As a whole, the assemblage reflects sedimentation of lacustrine environments in a moderately warm climate. No analogs of this assemblage are known in Primor'e. Its characteristic feature is a presence of great amounts of the curvate forms of old *Aulacoseira,* forming ring colonies, and of ellipsoidal *A. distans* var. *ovata.* Also found in the Miocene deposits in Oregon, the latter taxon was described by S.L. Van Landigham (1967) and noted as a guide form in the diatom assemblage found in the middle portion of the lower Miocene section in the Bes Konak basin, Turkey, and dated as older than 13 Ma (Servant–Vildary *et al.,* 1986). *Aulacoseira,* a similar oval was found by I.B. Tsoi in the interbeds of continental deposits among marine strata of the Yamato Rise, Sea of Japan, which were assumed to be of early Miocene age (Tsoi *et al.,* 1985).

We refer this diatom assemblage, clearly not reflecting the Miocene optimum, to the pre-optimum time, more exactly, to the earliest Miocene. This is confirmed by an absence of the *Actinocyclus* forms, including fresh-water diatoms of the *Coscinodiscus* (= *Pontodiscus*) genus, reliably known only since the late Burdigalian (Krebs *et al.,* 1987).

The date of 25.3 ± 1.6 Ma was obtained for a bed of volcanic sand at a depth of 36 m. If this date were reliable, the lower tephroid strata should correspond to the upper Oligocene. However, since such an old date is unique and is, to a certain extent, at variance with paleobotanic data, we accept it rather tentatively. The lower portion of the section at the west margin of Razdol'noe village (outcrop 9142) is a possible analog to the strata penetrated by borehole 44. The diatom assemblage recovered from it is very similar to that from borehole 44 and is likewise dated as the early Miocene (early Burdigalian).

Sediments, corresponding to the later stage of the early Miocene, were penetrated by borehole 20. The spore-and-pollen spectra recovered from them are characterized by an increased content of angiosperm pollen with respect to gymnosperm pollen and spores. The gymnosperm pollen group retains the spectrum type characteristic of the above-mentioned assemblages, whereas that of the angiosperm group changes. The Juglandaceae pollen became more prominent: *Pterocarya* (up to 4.9%), *Juglans* (up to 7.8%), *Carya* (up to 1.5%), and *Engelhardtia* (solitary grains). The *Myricaceae* pollen appeared, and the group of subtropical plants (Moraceae, *Liquidambar, Rhus, Ilex, Reevesia, Nyssa,* and *Zelkova*) became more abundant and diverse. The content of Fagaceae pollen is lower, and alder is predominant among Betulaceae, indicating that the area became water-logged, which is also supported by the presence of *Nyssa* and an abundance of *Taxodiaceae.* Palynospectra on the whole are noticeably more thermophilic, though do not contain the pollen of plants which are indicators of Paleogene floras.

The diatom flora is also more diverse (Fig. 3). The coarse-frustule *Aulacoseira praegranulata* (Jouse) Moiss. dominate as before (up to 40%), but *A. praedistans* (Jousé) Moiss. + var. *curvata* are also abundant (up to 20%). The appearance of *A. italica* (Kuetz.) Simonsen var. *italica* (up to 10%), *A. areolata* Moiss. (up to 5%), *Paralia scabrosa* (Oestr.) Moiss., *P. arenaria* (Moore) Moiss., (up to 7%), *Actinocyclus* sp. cf. *Coscinodiscus miocenicus,* and *A.* sp. + *A.* sp. cf. *C. lobatusva sibiricus* (3 - 4%)[1] has been noted. The species of the *Tetracyclus, Eunotia, Achnantes, Navicula, Pinnularia,* and *Gomphonema* genera are more diverse.

The assemblage as a whole reflects the existence of a large lake with a variable depth of up to 10 m. Three peaks of the *Actinella brasiliensis* tropical species are

[1] The taxonomic position of the *Actinocyclus* genus species has yet to be defined more precisely.

seen clearly in the diagram (Fig. 3), indicating warm climatic stages alternated with coolings and shallowings of the basin, where brown coals were deposited. This case may reflect an initial stage of the Miocene climatic optimum. Judging from the second appearance of the *Actinocyclus* genus forms, the age of the deposits corresponds to the second half of the Burdigalian. This was confirmed by fission-track dating. Four dates, 18.8 ± 1.1, 19.7 ± 1.1, 20.2 ± 0.8, and 20.9 ± 1.1 Ma, were obtained for three volcanic sand beds found within the interval. A diatom assemblage, very similar to the one described above (borehole 20, one of the *Actinella brasiliensis* peaks), was recovered from the upper beds of section 9142. A considerable hiatus in the roof of this section prevented us from tracing a further tendency towards a taxonomic change in the diatom assemblages.

Sediments corresponding to the middle Miocene were penetrated by boreholes 12, 35, and 37 (Vozdvizhenka depression). They contain abundant pyroclastics, and their spore-and-pollen spectra have main parameters similar to the spectra from borehole 20. Characteristics of these spectra are: the same predominance of angiosperm pollen over gymnosperm pollen and spores, the leading role of *Pinus* subgen. *Haploxylon* in the gymnosperm pollen group (up to 18%), an insignificant presence of the pollen of dark coniferous plants, and very sharp variations in the content of Taxodiaceae pollen (from solitary grains to 22%). Alder dominates in the angiosperm group (up to 35.5%), while other Betulaceae are present in lesser amounts. Fagaceae are represented modestly: the *Fagus* pollen content varies from solitary grains to 5% and the solitary grains of the *Quercus* pollen are present only in some samples. Characteristic of the latter is a resemblance to the pollen of Mongolian oak. The *Quercus* species typical of the Paleogene (*Q. gracilis, Q. graciliformis, Q. forestdalensis,* and *Q. conferta*) were not found. The pollen of representatives of a thermophilic flora is present in greater amounts. For instance, the content of *Liquidambar* pollen in some samples is as much as 4.8%.

The deposits of the above-mentioned sections yielded a compact group of fission-track datings, encompassing an interval of 14.9 - 15.9 Ma, i.e. corresponding to the Langhian. The clearly younger middle Miocene beds according to our data, have not been found.

A transition to the upper Miocene beds has been noted in outcrops 9142 and 9143, which, although situated only 300 m from one another, exhibit sequences separated by a considerable stratigraphic hiatus. The previously mentioned upper beds of section 9142 correlate, according to their diatom assemblage, with the sequence of borehole 20. The gravel strata with lenticular interbeds of sand, including volcanic sand, and tuffaceous silt exposed in outcrop 9143 have been referred, according to a highly representative assemblage of leaf flora, to the Ust'-Suifun Formation, corresponding to the regional horizon of the same name (Pavlyutkin *et al.,* 1985). Two diatom assemblages were revealed by a detailed sampling of the gravel strata. The lower part of the sequence (21.4 - 34.5 m) contains abundant *Aulacoseira praegranulata* (Jousé) Moiss. + var. *curvata* (up to 67%) and *Actinella brasiliensis* Grun. (up to 22%). The species of the *Actinocyclus* genus are also present. Based on the abundance of tropical elements (more than 54%) and the absence of many boreal species of the *Navicula, Pinnularia,* and *Eunotia* genera, this assemblage should be regarded as the most thermophilic of those studied.

The assemblage recovered from the interval of 0 - 21.4 m is substantially different. The species of the *Actinella* and *Actinocyclus* genera disappeared. *Aulacoseira jouseana* Moiss. and *A. areolata* Moiss. are also absent. The first representatives of the *Mesodyction, Stephanodiscus,* and *Cyclostephanos* genera appeared. A considerable amount of new species were found in such genera as *Tetracyclus* (*T. lacustris* Ralf + var., *T. chudjakovii* Pushkar, *T. ellipticus* var. *lancea* (Ehr.) Hust. f. *elongata* Must.), *Fragilaria* (*F. bicapitata* var. *lineolata* Moiss., *F. nitzschioides* var. *kamtschatica* Lupik., *F. spinosa* Skv., *F. miocenica* var. *tetranoides* Moiss., *F. transylvanica*), *Eunotia* (*E. delicatula, E. majuscula* Moiss., *E. goretskyi* Churs., *E. nikolskiae* Moiss., *E. polyglophoides* Sheshuk.), *Achnanthes* (*A. lanceolata* (Bréb.) Grun., *A. lapidosa* var. *robusta* Moiss., *A. scutiformis* Moiss.), *Pinnularia* (*P. borealis* Ehr., *P. esox* Ehr., *P. major* (Kuetz.) Cl., *P. mesolepta* var. *stauroneiformis* Grun.), and *Gomphonema* (*G. miocenica* Moiss., *G. parvulum* Kuetz.). *Aulacoseira praedistans* (Jousé) Moiss. is dominant in the assemblage.

On the whole, a tropical-subtropical diatom assemblage was replaced by a boreal assemblage, the result of the beginning of the cold spell which characterizes the late Miocene. The general characteristics of the assemblage resemble those of the diatom flora from the Lower Erman Subformation and the Medvezhka Formation in Kamchatka (Dolmatova *et al.,* 1984), which is, in the opinion of E.G. Lupikina, an analog of the Suifun (Ust'-Suifun, according to present-day appraisals) Formation of southern Primor'e and of the Lower Beletsk Horizon of Belarus' (Moiseeva, 1974; Kuklina and Moiseeva, 1980; Khursevich and Loginova, 1980). The beds with *Aulacoseira praedistans* in the Lower Erman strata are overlain and underlain by marine deposits of the *Denticulopsis hustedtii* Zone (upper Miocene). These facts and conclusions correspond to the fission-track date of 10.8 ± 0.9 Ma obtained earlier from a volcanic sand bed corresponding to the base of the upper diatom assemblage in section 9143 (Pavlyutkin *et al.,* 1984). The date of 14.5 ± 0.9 Ma from the top of section 9143 can neither be considered reliable, nor used for interpreting the results of the study, since the sample contained a mixture of volcanic glasses, which differed in color, index of refraction, and size and density of tracks. Apparently, a considerable admixture of older material has distorted the result of the dating.

Studying the Miocene sequences, we expected to find the deposits of the Miocene climatic optimum,

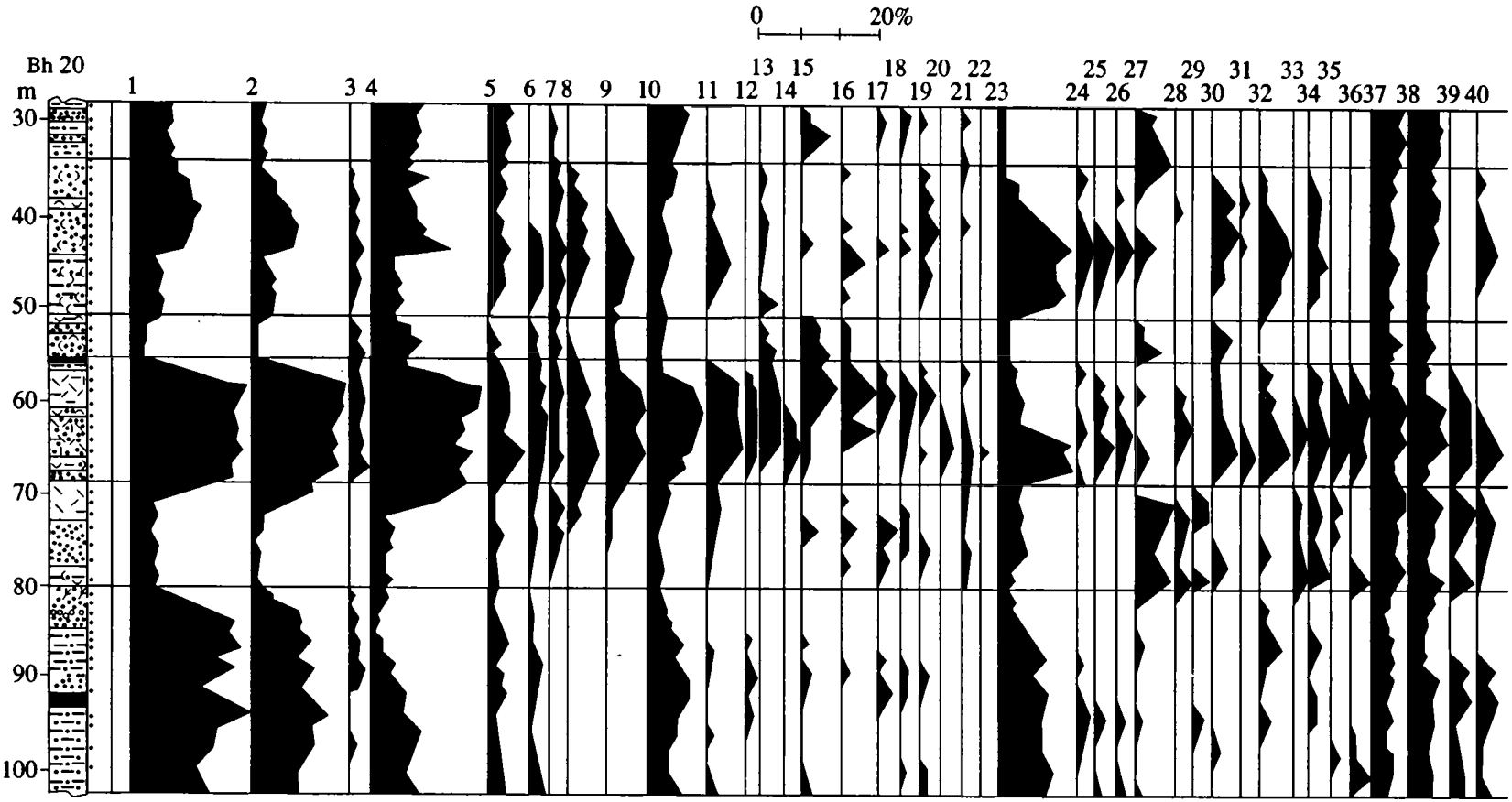

Fig. 3. Diatom diagram of the borehole 20 deposits.

1 – *Aulacoseira praegranulata*; 2 – *A. praegranulata* f. *curvata*; 3 – *A. praegranulata* var. *proxima* f. *curvata*; 4 – *A. praedistans*; 5 – *A. praedistans* f. *curvata*; 6 – *A. distans* var. *ovata*; 7 – *A. praeislandica*; 8 – *Paralia scabrosa*; 9 – *P. arenaria*; 10 – *Aulac oseira italica*; 11 – *A. areolata*; 12 – *A.* aff. *jouseana*; 13 – *Actinocyclus* cf. *Coscinodiscus miocenicus*; 14 – *A.* cf. *C. lobatus* var. *sibiricus*; 15 – *Tetracyclus lacustris*; 16 – *T. lacustris* var. *elongatus*; 17 – *T. lacustris* var. *latissimus*; 18 – *T. lacustris* var. *strumosus*; 19 – *T. ellipticus*; 20 – *Diatoma costatum*; 21 – *Fragilaria araulensis*; 22 – *F. nitida* var. *nitida*; 23 – *Actinella brasiliensis*; 24 – *A. punctata*; 25 – *A.* sp.; 26 – *Desmogonium* sp.; 27 – *Eunotia praerupta*; 28 – *E. clevei*; 29 – *E.* sp.; 30 – *Achnanthes miocenica*; 31 – *A. pinnata* f. *robusta*; 32 – *Biddulphia laevis*; 33 – *Stauroneis* sp.; 34 – *Navicula americana*; 35 – *N. jimboi*; 36 – *N.* sp.; 37 – *Pinnularia mesolepta* var. *stauroneiformis*; 38 – *P.* sp.; 39 – *Gomphonema miocenica*; 40 – *G. intricatum* var. *fossilis*.

which is clearly manifested in the sections of Northeastern Asia, Kamchatka (Akhmet'ev *et al.*, 1984; Laukhin *et al.*, 1984), and North America. A considerable warming in the middle Miocene in these regions following paleobotanical data, has been noted. It is manifested in an abundance of Taxodiaceae and Fagaceaea pollen, characteristic of the so-called "Fagus" or "Beech" Horizon. We were unable to reveal a distinctly pronounced climatic optimum in the sections considered above. This may be related to the fragmentary character of the obtained palynological data (many samples contained neither pollen nor spores). The "Fagaceae" beds at the west coast of Lake Khanka are recommended for placement at the level of transition from the middle to upper Miocene (Resheniya..., 1982). Moreover, the high content of beech pollen in palynospectra from these beds (and of its leaf imprints in taphofloras) is combined with a predominance of the dark coniferous plants, representatives of the *Picea, Tsuga,* and *Abies* genera. This type of spectra demonstrates that cooling and humidification of the climate occurred, resulting in the lowering of the beech forest tree line.

Thus, we can make the following conclusions:

(1) The stratotype sections of the Tavrichanka and Khanka horizons are of Paleogene age (without specifying) and may not be assumed to be a regional stratigraphic equivalent of the lower-middle Miocene, at least in the absence of additional studies and the coordinated results of different paleobotanical methods.

(2) Radiometric (fission-track) data, obtained independently of paleobotanical investigations, in combination with the latter suggests that the tephroid strata, having distinct lithostratigraphic boundaries, represent an equivalent of the lower-middle Miocene. The base of the strata might possibly correspond to a geochronological level somewhat lower than the Paleogene–Neogene boundary. The age of the top of the tephroid strata has not been determined.

(3) No traces of a distinctly pronounced climatic optimum have been found by a palynological analysis of the studied strata. A diatom analysis indicates several equivalent levels with thermophilic assemblages. Apparently, only further in depth investigations may lead to a more definite resolution of this problem.

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Late Cenozoic Paleoclimatic Events in the Arctic Ocean

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Abstract – Materials on the most important late Cenozoic paleoclimatic events in the Arctic Ocean are presented. Based on data obtained, the highly dynamic changes of environments are discussed.

Oceanic heat transfer in the late Cenozoic was one of the important factors causing the paleoclimatic changes in high and temperate latitudes in the northern hemisphere. A global heat transfer in high latitudes is clearly dependent on the water circulation, or mass water structures, the global ice volume, and sheet glaciation. Many researchers have shown that arctic cold waters, enriched with oxygen, ventilate the deep waters of the North Atlantic, affect food resources and carbon cycles, and also have an impact on the mechanism of global heat transfer and temperature distribution both in oceans and the atmosphere. Therefore, the most important problem to be addressed in our paper is to define the time interval in which the deep arctic waters appeared in the North Atlantic. The problem of initial glaciation in the Arctic Ocean should be considered as well.

The available materials on glaciation in the Arctic region admit a wide time range, from 30 to 0.9 Ma (Clark, 1977; Clark *et al.*, 1980; Herman, 1984; Thomas and Thide, 1991).

According to data on the Eurasian shelf seas of the central and eastern Arctic region, the ice appeared in the Arctic Ocean in the Messinian, while at the same time the cryolithozone and paleotundra came into existence on the adjacent land (Alekseev *et al.*, 1991; Zarkhidze *et al.*, 1991). Recent data on the ice-rafted material in sediments of the Norway-Greenland sedimentary basin consistently determined the age of deposition to be earliest late Miocene. This data is the first evidence substantiated geologically, and allows us to infer that at least 10 Ma ago ice formed in the Arctic Ocean, and, in contrast to the Paleogene, a cooling in the northern hemisphere resulted in glaciation in the Arctic region.

It is interesting to consider when the regular connection between the Arctic and Atlantic oceans appeared, i.e., when was the isolated Arctic basin incorporated into the world ocean system. Of interest in this respect, the investigations of R. Vogt (1986) elaborate a geodynamic model of destruction (the result of sea-floor spreading) of the threshold separating the Arctic and Atlantic oceans. According to his data, a regular connection between these two oceans started during the paleomagnetic epoch 13, about 30 - 35 Ma, due to the westward motion of Spitsbergen and the formation of the Fram Strait gateway. But this has not been proven definitively. Paleomagnetic investigations made by Norwegian researchers on the Yermak Plateau suggested that the regular connection between the Arctic and Atlantic oceans set in, more than likely, at the middle-late Miocene boundary. Since that time, the arctic deep waters have found their way into the Atlantic ocean (Kristoffersen, 1990).

According to the material available, the Arctic and Pacific oceans became connected in the late Miocene (10 to 8 Ma ago). This connection was repeatedly interrupted throughout the late Cenozoic (Laukhin and Patyk-Kara, 1991).

Investigations in the Greenland-Norwegian basin established that maximum concentrations of the ice-rafted material exist in the upper Miocene–Pliocene deposits. These concentrations are confined to the following time intervals: 9.5 - 7.0; 5.7 - 5.4; 4.1 - 3.9, and 3.5 - 3.2 Ma. These intervals correspond to periods of cooling in the northern hemisphere (Thomas and Thide, 1991). The first of these intervals is distinguished by its long duration, about 1.5 Ma, perhaps the result of insufficient investigation. On the other hand, such a long period of cooling at the beginning of the late Miocene may indicate an initial stage of dramatic change in paleoclimatic conditions in the subarctic and temperate zones of the northern hemisphere.

Subsequent cold intervals lasted 200 - 300 ka, whereas, with the exception of the earliest one, which lasted 1.5 Ma, the warm periods were as long as 500 - 600 ka. Thus, in the Greenland-Norwegian subarctic basin, distinct climate cycles are established. The time interval of 9.5 - 7.0 Ma corresponds to the opening of the Fram Strait during the course of sea-floor spreading. Arctic waters found their way through this gateway southward giving rise to the East Greenland cold current. The period of cooling 5.7 - 5.4 Ma correlates with the Messinian crisis, and the period 4.1 - 3.9 Ma corresponds to the formation

time of the Istmuth of Panama. These concepts allow us to relate periods of cooling in the late Miocene–Pliocene with the geodynamic events that caused changes in the water mass structure, which in turn resulted in the onset of new oceanic currents and in a periodic drift of the arctic ice into the Greenland-Norwegian basin. Nevertheless, although corresponding to Pleistocene glacial epochs in duration, the range of cold intervals was found to be much less when compared to that of the warm intervals. Studying the late Cenozoic flora of Iceland, M.A. Akhmet'ev (1978) reported advanced cooling since the second half of the late Miocene. He conceives that a decrease in average annual temperatures of 5 to 6°C occurred at that time, with the average temperature of the coldest month as low as –15°C and the absolute minimum as low as –35 - –40°C.

At the end of the middle and with the beginning of late Miocene, the vegetation in Iceland, was recorded to be depleted in broad-leaved species as a result of cooling; although the Moculdalur time (9.5 - 8.5 Ma) registered a warm episode (Akhmet'ev, 1978). Thus, a discrepancy between the paleobotanic data and the material related to the dating of the first late Miocene cooling in the Greenland-Norwegian basin is evident. Subsequent investigations are needed to clarify this question, though the ice-rafted material deposited at the beginning of the late Miocene has already been found at several sites of deep sea drilling.

The origins of tundra, and probably of permafrost as well, are related to substantial climatic cooling at the beginning of the late Miocene on the Chukchi Peninsula (Laukhin and Patyk-Kara, 1991). Although, J. Matthews (1976) does not exclude the appearance of both events in the Canadian Arctic already in the middle Miocene.

Advanced cooling in the late Pliocene and Pleistocene resulted in repeated ice sheet glaciations. Since that time, a new stage in the earth's geological history, and, hence in paleoclimates, commenced. This raises a question concerning the onset time of the ice sheet glaciation which should be discussed. According to recent publications, the first onset of sheet glaciation in Iceland occurred 3.1 Ma (Eiriksson and Geirsdotter, 1991). Six horizons of tillites are distinguished in the time interval of 3.1 - 2.5 Ma, and ten horizons in the 2.5 - 0.7 Ma interval. The abundance of tillite horizons should, first of all, indicate highly dynamic glaciation, as well as a series of critical climatic changes. However, these numerous horizons may also be related to the regional position of Iceland, as well as to its volcanic activity. Subsequent volcanic eruptions could cause a partial destruction of ice sheets which, after the eruption cesation and lava cooling, could form in the same area again, initiating the formation of tillite horizons of the same geological age, but occurring at different levels. It should be noted that the previously age of 3.1 Ma refered to the Borgarfjordur section located in central Iceland. However, the tillite horizons in the northern part of the Island and its shelf (the sections of the Tjornes Peninsula and Flatey Island, respectively) are dated not older than 2.5 Ma. Thus, the last data pinpoints the onset time of the ice sheet in Iceland.

Recent investigations in the Norwegian Sea clarified the age of the initial stage of the Scandinavian glaciation. A sedimentary sequence found in the Norwegian Sea establishes the Scandinavian origin of the characteristic ice-rafted material and implies that the onset of the Scandinavian glaciation took place 2.56 Ma ago, opposed to the 2.4 Ma age adopted earlier. This value correlates well with the age data on the first glacial phase in the western part of the Arctic Ocean (Haflidason *et al.*, 1991; Hollemann *et al.*, 1992; Jensen *et al.*, 1988).

The oldest moraine on the mid-Norwegian shelf dates back to 1.1 Ma. Besides the older sequence of glaciomarine sediments, two glacial levels separated by interglacial marine deposits are recorded in the early Pleistocene strata (from 1.1 to 0.7 Ma). In the middle Pleistocene, three glacial phases of the Scandinavian glaciation are distinguished (Haflidason *et al.*, 1991).

Materials presently available allow us to date the beginning of glacial history of Greenland back to 2.4 Ma, though some researchers argue that this event was initiated 3.0 to 3.5 Ma (Zarkhidze, 1991; Zarkhidze *et al.*, 1992).

Therefore, the advanced cooling promoted by the ice sheet onset dates back to 2.5 - 2.4 Ma, at least in the subarctic region and, later on, on the arctic plains occupied by shelf seas (excluding the East Siberian and Laptev seas).

In northeastern Chukchi Peninsula, two moraines left by the mountain–valley glacier were revealed and dated as 3.0 and 2.5 Ma old, respectively (Laukhin and Patyk-Kara, 1991). Within the arctic plains occupied at present by the East Siberian and Laptev seas, the earliest traces of permafrost seem to be confined to the same time interval. However, even in the Arctic Ocean, the climate was not always similar to that characteristic of the glacial epoch. Recent investigations of the Arctic Ocean within the Gakkel Ridge have shown that complex changes in sedimentation processes, which were caused by climatic conditions, occurred in this aquatorium during the past 250 ka. The interval of 250 - 128 ka is characterized by glacial and glaciomarine sedimentation, and within this interval in the period of 186 - 128 ka at least three short-term warmer climate phases are recognized. During the last interglacial epoch and in the Holocene climatic optimum, the climate was considerably warmer, and, according to data from studies of the Lomonosov Ridge, only seasonal sea ice formed during these periods (Nowaczyk *et al.*, 1992). During the

Holocene climatic optimum, no glaciomarine sedimentation occurred in the Barents Sea, and during the winter time the sea ice zone was believed to be limited. At the same period, ice domes are assumed to have disappeared in the Franz Josef Land (Lavrushin *et al.*, 1990).

Intensive dynamics of climatic and environmental changes were also discovered in the Greenland and Norwegian seas. They were caused by fluctuating spatial and temporal distributions of various types of surface waters. Transitional glacial and interglacial stages are recorded at the boundaries of oxygen isotope stages 14 to 13, 12 to 11, and 6 to 5 (Osterman, 1992). At those time intervals corresponding to isotope stages 11.1, 5.1, and 1.0, conditions were similar to those of the interglacial epoch due to Atlantic water invasion. Short-term penetrations of Atlantic waters are recorded for isotopic stages 9.3, 8.6 to 8.5, 7.5, 7.3 to 7.1, 6.5, 5.4, 5.3, and 3.1. Isotopic stages 12, 10, 8.4, 7.4, and 5.2 correspond to cold climatic environments of glacial and interstadial stages (Henrich *et al.*, 1992). For some of them, a relative decrease in surface temperatures amounting to 8 - 10°C was recorded in the North Atlantic (Ruddiman *et al.*, 1986; Jung *et al.*, 1992).

A spatial distribution of warm Atlantic waters resulted in differing intensities of sea ice melting. Data on the Norwegian-Greenland basin illustrates this well. In its eastern part, the zone of the paleo-Gulf Stream, a great amount of ice melt water was registered in the time interval corresponding to isotopic stages 1, 5.1, and 5.5. In the northern Greenland Sea and in the Fram Strait, the ice melted considerably slower during the same stages (Baumann, 1992). Thus, we suggest that the climatic changes in the Arctic region were regional and peculiar in character.

Investigations in the Fram Strait allowed us to date the glacial events on Spitsbergen as 130 - 140, 112 - 110, 75 - 50, and 20 - 10 ka.

Studying the ESP-chronostratigraphy of marine sediments of East Eurasia revealed stages of high sea level which correlate to interglacial and interstadial events: 550 and 450 - 385 ka (Kotel'ny Island, New Siberian Islands), 300 ka (Severnaya Zemlya), 170 ka (Western Siberia), 120 - 82 ka (White Sea, Arctic Islands), and 68 - 52 ka (Taimyr) (Molod'kov *et al.*, 1992). Detailed chronostratigraphic investigations of the Severnaya Zemlya Archipelago revealed that during the past 300 ka the augmentation of ice sheets occurred in the intervals of 200 - 300, 120 - 170, 70 - 80, and 12 - 19 ka (Makeev *et al.*, 1992).

Clearly, the position of the polar front affected the climate of the Arctic region. Investigations in the northeastern part of the Norwegian Sea confirmed intensive dynamics and large-scale changes in the polar front position during the past 50 ka. During isotopic stage 3, an open water basin existed here, and at the boundary of stages 3 and 2, the polar front, as well as a zone of arctic and polar water mixing were located in this area. In the time interval of 20 - 15 ka, the northern part of the Norwegian Sea was free of ice (Dokken, 1992).

In the time interval between 15 and 11 ka, unstable climatic conditions related to climatic changes were recorded in the Arctic Ocean and adjacent seas. It has been established that warmer conditions occurred at 14.6, 14.2, and 13.2 ka (Karpuz, 1992). Minimum temperatures of the surface water in the Bölling–Alleröed period are recorded for the time intervals of 12.9 - 12.8, 12.5 - 12.4, 12.3 - 12.0, 11.9 - 11.5 ka (Karpuz *et al.*, 1992; Karpuz and Jansen, 1992; Wienelt *et al.*, 1992).

Arctic conditions existed during the late Dryas in the Greenland-Norwegian basin.

The Holocene is also characterized by climate variability. We have already mentioned its climatic optimum. Isotopic and biostratigraphic investigations in the North Atlantic revealed that the cooling of surface water occurred 8.5, 5.8, and 1.4 ka. In addition, we should point out that on the coast of the Laptev Sea and on the Taimyr shore the January and July temperatures are thought to have increased by 2 - 4°C at 6.0 - 5.0 ka (Bobrov *et al.*, 1987).

For the past 1000 years, valuable data on the limits of glaciers was obtained in the North Spitsbergen. Five short-term phases of glacier advance were recognized for this interval (Furrer, 1992).

Thus, the review of the material available for the Arctic basin suggests diverse climatic changes. Two stages, the late Miocene–Pliocene and the glacial stage after 2.5 Ma, should draw our attention. Nevertheless, ice appeared in the Arctic Ocean before the late Miocene. The stage of climatic changes in the Arctic Ocean generally considered as "glacial" was characterized by dramatic changes in climate. These changes allow us to reconsider the existing concept concerning its conservation.

Climatic changes in Arctic regions were the result of many events also affecting global oceanic heat transfer. First, this is related to geodynamics, changes in the Earth's orbit parameters, evolution of sea current structure, etc. In this respect, the more intensive dynamics of climatic changes in the western sector of the Arctic region should be noted, because the Fram Strait remained free of gletcher ice during the glacial periods and the Norwegian-Greenland basin was an area where waters of the Arctic and Atlantic oceans mixed often.

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Evolution of the biosphere

Chronology of Late Cenozoic Biotic and Abiotic Events in Western Siberia

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Abstract – The geochronologic succession of currently revealed biotic and abiotic events of the late Cenozoic in western Siberia is considered. The results show that fundamental changes in the biota and sedimentation occurred simultaneously with global climatic changes.

The general regularities of faunal and floral evolution in Siberia during the late Cenozoic are well known. Nevertheless, we still lack a precise knowledge of (1) the chronology of biotic and abiotic events, (2) the regularities of biotic evolutionary trends, (3) the causes of biotic reorganization, and (4) the extinction, mass migration, and adaptation to rapid changes in the natural environment. These key regularities are necessary to predict scenarios of the near future.

Nearly complete successions of sediments, small mammalian fauna, and fresh-water molluscs were revealed for the upper Miocene and the Pliocene in the southern West-Siberian plain. The main stages of fauna evolution were outlined, and the summary magnetostratigraphic section was compiled. A detailed and well-justified stratigraphic scheme (the table) was worked out. The data obtained on sedimentation environments and the history of the biota allowed us to trace the main climatic events. Their chronology was established using paleomagnetic and biostratigraphic data, and a correlation with climatic changes in other regions was defined (Zazhigin and Zykin, 1984; Zykin, 1982, 1988, 1991; Zykin *et al.*, 1989, 1991). The composite sequence of the West Siberian Pliocene is one of the most complete continental Pliocene sections (especially for the warmest lower period) with regard to the degree of detail in the subdivisions and the representation of geologic, paleobiologic, and paleoclimatic events. This allows us to use it as a stratotype for the terminal Miocene and the Pliocene of Central Asia as a whole.

Unfortunately, palynological data on the upper Miocene and the Pliocene, obtained mainly in the 1970s (Volkova, 1971, 1977; Volkova and Panova, 1975, and others), is limited and poorly correlated to the present stratigraphic subdivisions (Zykin *et al.*, 1989, 1991). Many of these subdivisions are not characterized by paleobotanic data, which prevents a comparative palynological analysis and spatial reconstruction of vegetation for all presently established time intervals in the late Cenozoic. The correction of palynological data in accordance with the results of the detailed stratigraphic works mentioned above will be an object of further research. Now, in order to avoid distorting the data, we must regard them separately, according to our stratigraphic concept in the above cited works.

At the end of the late Miocene (Pavlodar Time) the climate reached its most arid stage. This stage is evident in such events as the formation of the subaerial, red-carbonate-clay cover filling the river valleys and lake depressions; the reduction of river outflow; the extinction of large groups of mammalian fauna; and the expansion of the desert zone (Zykin, 1982, 1991; Zazhigin and Zykin, 1984; Zykin *et al.*, 1989, 1991). The Pavlodar Arid Phase, clearly traced in the interior of Asia, corresponds, according to paleomagnetic and biostratigraphic data (Zykin *et al.*, 1989, 1991), to the global decrease in the ocean level by 50 - 70 m at the end of the Miocene and to the associated Messinian salinity crisis (Adams *et al.*, 1980; Hsü *et al.*, 1978), which caused significant continentalization and an aridization of the climate.

At the beginning of the Pavlodar time, the flora was significantly enriched by forms of grass-shrub vegetation, and the existence of assemblages including *Artemisia,* Poaceae, Chenopodiaceae, Fabaceae, *Brassicaceae*, and *Ephedra*, with a reduced proportion of tree species (Volkova, 1977). The broad-leaved *Ulmus*, *Quercus*, and *Corylus* were members of this ecosystem. This vegetation was of the forest-steppe type with a large participation of xerophytes. The aridization of the climate along with the progression of desert zones

Stratigraphic scheme of the Pliocene and the Eopleistocene in the southern West Siberian plain (climate: C – cold, W – warm)

System	Group	Series	Ma	Paleomagnetic Scale		Regional Stratigraphic Subdivisions					Local Stratigraphic Subdivisions				Climate
						Regional Stage	Horizon	Typical Complexes: Mammals	Molluscs: fresh-water	Molluscs: terrestrial	Kuibyshevo–Ishim area	Petropavlovsk–Ishim area	Pavlodar–Irtysh area	Omsk–Irtysh area	C W
Quaternary	Eopleistocene		1		Matuyama		Kochki	Razdol'ino	Karagash						
								Kizikhino			Karagash Formation				
Neogene	Pliocene	upper	2			Kokchetav	Irtysh	Lebyazh'e	Mukkur		Mukkur Formation			Mukkur Formation	
									Lebyazh'e				Irtysh Formation		
			3		Gauss		Selety	Kyzyl-Aigir			Selety Formation		Aksor Beds Selety Formation		
		lower				Irtysh	Biteke	Andreevo				Liveno Formation	Liveno Formation	Liveno Formation	
			4		Gilbert			Biteke			Biteke Formation			Biteke Formation	
							Novaya Stanitsa	Krutaya Gorka						Krutaya Gorka Formation	
								Peshnevo				Peshnevo Formation		Peshnevo Formation	
			5					Cherlak			Kuskol Formation	Rytovka Formation	Rytovka Formation	Rytovka Formation	
	Miocene	upper			5			Novaya Stanitsa				Novaya Stanitsa Formation		Novaya Stanitsa Formation	
			6		6		Pavlodar	Pavlodar			Pavlodar Formation		Pavlodar Formation		

were typical of the second half of the Pavlodar Time, which is marked by red deposits. The trees disappeared completely; an intensive enrichment of the vegetation assemblages by xerophytes of the Asteraceae family and *Ephedra* took place (Volkova and Kul'kova, 1988).

Cooling phase and humidization of the climate occurred at the Novaya Stanitsa Time (about 5.1 - 4.9 Ma on the paleomagnetic scale) as evidenced by the wide distribution of limnic sedimentation, the appearance of large amounts of modern Palaeoarctic species of fresh-water molluscs (up to 45%) against the background of thermophilic Sino-Indian malacofauna, and the appearance of hydrophylous mollusc forms among the terrestrial molluscs. According to climate-stratigraphic data, these events correspond to the Lago-Mare stage of the terminal Messinian (Zykin, 1991).

The semidesert vegetation was substituted by forest-steppe formations. *Plicea, Abies, Tsuga* of boreal-taiga formations, broad-leaved *Ulmus* and *Quercus* and, small-leaved *Betula* and *Alnus* forms occurred in forests preserved along river valleys. The riparian forms of the *Myriophyllum, Nuphar,* and *Polygonum* genera prevailed among grass-shrub vegetation. The composition of the flora indicates both humidization and some cooling of the climate, though it was still warm. The vegetation was similar to that of the forest-steppes of Manchuria (Volkova and Kul'kova, 1988).

The early Pliocene – from 4.9 Ma until 3.2 Ma – was characterized by a warm and relatively stable climate, which contributed to the wide dispersal of thermophilic, Sino-Indian, fresh-water malacofauna and Ruscinian mammals on the southern plains. Periods of relative humidity alternated with arid phases. This is clearly reflected in the cyclic structure of continental sedimentation, the cycles beginning with aqual limnic or alluvial deposits and finishing with subaerial deposits, which show horizons of large carbonate concretions. Five cycles of humidization (Zykin *et al.*, 1989) are presently recognized in the lower Pliocene. Fluctuations in the thermal regime were generally of a relatively low amplitude against the background of the gradual cooling. The Pliocene climatic optimum in the southern plain corresponds to the beginning of the Pliocene, the Rytovka–Peshnevo Time – 4.9 - 4.4 Ma (Zykin, 1991). This is apparent in the reduced abundance of Palaeoarctic forms of fresh-water malacofauna, in the wide dispersal of Sino-Indian species, and in the formation of non-carbonate, red limnic deposits.

An insignificant cooling took place during the Krutaya Gorka, typified by the impoverishment of fresh-water malacofauna, the increase of Palaeoarctic species (up to 53%), and the extinction of some thermophilic, Sino-Indian and West-Siberian genera (Zazhigin and Zykin, 1984).

The Biteke–Liveno Time (4 - 3.4 Ma) was again characterized by a warming of the climate (Zykin, 1991; Zykin *et al.*, 1991). Thermophilic, Sino-Indian and West-Siberian elements once again prevailed among the fresh-water malacofauna, and the Palaeoarctic elements were reduced, though their amount is slightly larger than that at the Pliocene optimum.

A sharp reorganization of the biota, connected with a significant cooling and aridization of the climate (Zykin, 1991; Zykin *et al.*, 1989, 1991), occurred in the south of the West Siberian plain at the beginning of the late Pliocene – about 3.2 Ma (the beginning of the Mammoth Episode). Almost all thermophilic species of fresh-water molluscs died out, the malacofauna acquired Palaeoarctic features, xerophyte species of terrestrial molluscs appeared, and the Ruscinian forms were replaced by Villafranchian mammalian fauna. This climatic event in the southern plain coincides with the extinction of fresh-water malacofauna in central Japan at 3.1 Ma (Matsuoka, 1987), and it corresponds to the turning point in the global climate evolution, which is reflected in the oxygen isotope curve for the sequence V 28-179 at 3.2 Ma (Shackleton and Opdyke, 1977).

A significant cooling of the climate in the southern West Siberian plain again took place 2.5 Ma, at the beginning of the Irtysh. This time, the malacofauna acquired a completely modern form, and all thermophilic elements died out. The climate now became more like the modern climate (Zazhigin and Zykin, 1984). This event coincides with the Pretiglian in northwestern Europe, and corresponds to the cooling of the climate, which is clearly traced throughout Eurasia, North America, and the southern hemisphere.

The next warming of the climate corresponds to the Tiglian episode in Europe and is recognizable due to the appearance of thermophilic, Middle Asian malacofauna species, which occurred at 2.1 Ma – the end of the late Pliocene (Mukkur Time).

Fluctuations in the climate from warm and humid to cool and dry periods are also apparent in the development of the Pliocene flora: the abundance of broad-leaved plants was extremely low and the flora enrichment in modern West Siberian species of the boreal-taiga (and later in subarctic and arctic elements) was occurring gradually.

Trees and riparian plants were completely absent in Rytovka Time. The flora composition was extremely poor and consisted exclusively of xerophytic grass-shrub groups composed of *Artemisia, Aster, Ephedra,* Poaceae, Chenopodiaceae, and Brassicaceae, all of which points to the development of semidesert and desert landscapes, which, in the Pliocene, occupied a territory that today is a steppe or forest-steppe zone.

Later, palynoflora of the Krutaya Gorka Time was represented by West Siberian genera. Its main feature was the almost entire absence of the pollen of woody plants. Single grains belong to *Betula, Alnus, Pinus silvestris,* and *P. sibirica*; broad-leaved plants did not occur. Gramineae constitute the greater part of the assemblage (70 - 80%), in which the xerophytes Asteraceae, Chenopodiaceae, Brassicaceae, Umbelliferae, and *Thalietrum* follow. This palynoflora reflects a new stage in the development of steppe vegetation (with birch groves). The landscape was similar to the

modern steppes of the southern part of western Siberia and indicates cooling and humidization of the climate (Volkova and Kul'kova, 1988).

According to palynological data, the Biteke Time (4.0 - 3.7 - 3.3 Ma) is related to the thermal optimum of the Pliocene. In climatic aspect, it was extremely unhomogeneous. The flora was 80% of grass-shrub vegetation (*Artemisia*, *Poaceae*, *Chenopodiaceae*, and *Ephedra*) during periods of aridization, and was enriched in tree species during episodes of humidization. The birch was the main type of tree flora. Species of the broad-leaved genera *Tilia*, *Ulmus*, *Quercus*, *Carya*, *Platycarya*, and *Fagus* grew along river valleys.

The following phases were established in the development of the Biteke vegetation: steppes with the herbage–Artemisia–Chenopodium assemblage, forest-steppe assemblages that included the broad-leaved *Ulmus*, *Fagus*, *Juglans*, *Quercus*, *Carya*, and Castaneae, steppe-semidesert vegetation, and forest-steppe flora dominated by grass and some broad-leaved species. The climate was at that time significantly warmer than at present.

The forest-steppe vegetation persisted to the end of the early Pliocene (about 3.2 Ma); the dominant forms among the grass-shrub plants were species of the *Chenopodium* genus.

During the Podpusk–Lebyazh'e Time [Irtysh Time after Zykin (1991)], the flora was represented by modern West Siberian species. It reflects a succession of three types of vegetation, namely forest-steppe with random broad-leaved *Corylus* and *Quercus*, steppe vegetation dominated of herbage and *Chenopodium* forms; and again the forest-steppe assemblage, which now lacks broad-leaved plants. The absence of the latter from the flora composition, along with the increase of grass forms, indicate the aridization and gradual cooling of the climate. As a result, the forest-steppe vegetation in southern West Siberia, preserving its general form at the end of the Barnaul Time, had separate elements of Arctic flora: *Betula nana* and *Lycopodium alpinum*.

Palynoflora from the Karagash and Mukkur Formations (2 - 1.87 Ma) has not been well studied, and it is impossible to speculate on the character of corresponding landscapes.

The Eopleistocene chronostratigraphy has not been studied well enough. Poorly correlated beds, units, and formations in various regions of West Siberia with limited data on their stratigraphic volume, climatostratigraphy, and detailed subdivision into horizons, are attributed to this time interval. For this reason, the reconstruction of biotic and abiotic events may be discussed only in a very broad sense. It is known that limnic sedimentation occurred widely in the southern West Siberian plain during the Eopleistocene (Kochki Time). It is assumed that there were numerous changes in vegetation types, such as a substitution of forest-steppe landscapes by dark-coniferous forests and herbage–*Artemisia–Chemopodium* steppes by green-moss swamps (Volkova, 1991).

Two types of vegetation existed during the second half of the Kochki Stage (Ubinskaya-Erestnaya Time) in southern West Siberia. At the beginning, there were areas of green moss swamps combined with steppes covered by cryophilic assemblages of Graminiae, *Artemisia*, and *Chenopoduim* in combination with elements of subarctic flora. Later, this flora was replaced by forest-tundra assemblages with widely occurring spruce, dwarf birch, and arctic *Lycopodium*. This type of vegetation may be considered typical of the post-glacial stages (Volkova, 1977).

The chronostratigraphic scale of the Pleistocene, the main subdivisions and boundaries of which are justified by radiometric ages and correlated with the oceanic oxygen-isotopic scale, was suggested by Arkhipov (1987, 1989, 1991). A time-space alternation of two flora types was recognized in this period (Volkova, 1977).

The first type was characterized by arctic and subarctic elements in combination with species of the Graminae family and xerophytes. Flora composition indicates the development of tundra and forest-tundra landscapes to the south of the glaciation boundary. The flora contains large amounts of meadow-swamp species in combination with xerophytes and arctic *Lycopodium*, characteristic of the tundra hyperzone, which was spread over all the territory of West Siberia south of the glacier during the last glaciation period (Sartan Time). The southward migration of landscape zones during the Sartan Time is estimated to be 1100 km. This distinguishes it from earlier (700 - 800 km) glaciation stages (Volkova, 1971, 1977).

Flora of the second type consisted of boreal-taiga genera and species developed during the interglacial stages. The northward migration of the forest zone boundary was as much as 500 - 600 km. During some epochs (Kazantseva Time), this boundary apparently coincided with the coastal line of the arctic seas. A complex pattern of landscape evolution was peculiar to the Holocene. The cold climate persisted during the pre-Boreal and the beginning of the Boreal periods. Tundra and forest-tundra landscapes were located south of 61°N latitude at the beginning of the Boreal period (9.0 ka). The expansion of forest vegetation occurred during the second half of Boreal time. The formation of dark coniferous forests with elements of broad-leaved flora coincides with the beginning of the Atlantic Stage (6.0 ka). The maximal propagation of broad-leaved flora occurred between the end of the Atlantic and the beginning of the Subboreal periods (5.5 - 4.5 ka). The formation of modern geographic zonation corresponds to the Subatlantic period (Volkova and Levina, 1982).

Finally, it is necessary to conclude that the main global biotic and climatic events are clearly recorded in the sequences of the upper Cenozoic continental deposits in the West-Siberian plain. In the southern region of Siberia, the primary changes in the evolution of the biota, the environment, and the late Cenozoic climate took place at the terminal Miocene (at the end of

the Pavlodar). They are marked by the extinction of many mammal groups and by a significant aridization of the climate. These events correspond in time to a global decrease in the ocean level, and the associated Messinian salinity crisis.

A distinct cooling occurred during the Novaya Stanitsa Time, when modern Palaeoarctic species, together with thermophilic Sino-Indian malacofauna, became widespread (up to 45%) for the first time in southern Siberia, and boreal-taiga vegetation propagated in semidesert landscapes. Since this time, the climate migration processes have played an increasingly important role. They were marked by numerous changes in the relationship between Sino-Indian and Palaeoarctic species of malacofauna in the early Pliocene, whereas the alternation of forest-steppe assemblages, containing insignificant amounts of broad-leaved plants with steppe-semidesert assemblages similar to that of the modern type, took place.

The two warmest intervals were recognized in the early Pliocene. The Rytovka–Peshnevo interval is the first, according to for the malacofauna; the second, according to palynology, is the Biteke interval. This contradiction needs to be resolved, which requires a comparative study of palynoflora from all presently distinguished stratigraphic levels of the Pliocene.

The most important climatic and biotic reorganization boundary, corresponding to the turning point in the development of the global climate, is 3.2 Ma, i.e., at the beginning of the Mammoth Episode of the Gauss Epoch. It is recorded in the sharp extinction of the Sino-Indian malacofauna, in the replacement of Ruscinian mammals by the Villafranchian fauna, and in the appearance of flora represented by modern West Siberian species.

The final formation of the modern fauna and flora composition, with the complete extinction of all thermophilic elements, apparently took place at 2.5 Ma.

Since the second half of the Pliocene, and particularly during the Quaternary, biota began to be affected by rapid (dozens to hundreds ka) natural processes, such as catastrophic coolings and warmings of the climate, numerous glaciations in high and middle latitudes, widely distributed permafrost, dusty hurricanes in periglacial zones, rapid drops and increases in the ocean level, terminations associated with a huge outflow of glacial melt water, etc. These global processes caused events such as the grand time-space migrations of fauna and the appearance of mixed tundra-steppe groups; the shifts of up to 1000 - 1100 km southward (during glaciations), and of up to 500 - 600 km northward (during interglacial periods) of vegetation zone boundaries and the resulting formation of glacial interglacial cycles; with alternations of two main types of vegetation: the specific periglacial and the modern vegation. Both the evolutional and the migrational-climatic regularities in biota changes are important for an ecological prognosis.

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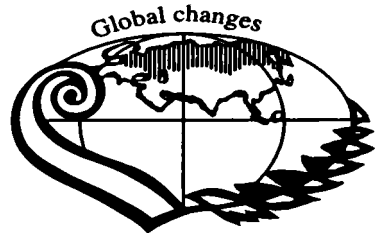

Evolution of the biosphere

Climatic Changes in Northeast Asia during the Pliocene–Pleistocene

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Abstract – During the last 5 Ma, some of the numerous climate fluctuations in Northeast Asia attained high amplitudes. The tundra and probably the permafrost first appeared as far back as 5 or 6 Ma on the lowlands of that area. In the lower reaches of the Kolyma River, the southern type of taiga last appeared about 4.5 - 4 Ma, and the northern type of taiga about 40 ka; the first cryoarid tundra-steppes appeared on these territories about 1 - 0.8 Ma. The mountain–valley glaciers first ventured deep into the lowlands of Northeast Asia (in the North of East Chukotka) 3.2 (3.4?) Ma. On the whole, the sequence of paleoclimatic events in the northern regions of Northeast Asia corresponds to climatic events in the Northern Hemisphere, and also has analogs in the Southern Hemisphere.

In the last few decades, the climate-stratigraphic method was increasingly used not only in Quaternary geology, but in Neogene, and particularly Pliocene studies. To justify the subdivision in local and regional stratigraphic scales, it is extremely important to know how widespread the paleoclimatic events used for climate-stratigraphic interpretations were. The correlative value of paleoclimatic events, both inter-regional and sub-global, is of supreme significance. Merely local climatic changes are also important for detailed local stratigraphic scales. Until quite recently, it was believed that the northern regions of Northeast Asia, at least since the second half of the Pliocene, had an extremely severe climate, which changed only slightly in the middle of the Pliocene. Therefore, the paleoclimatic reconstructions of the Pliocene and Pleistocene for these areas that in fact show considerable and sudden fluctuations of climate may be essential for regional climate-stratigraphic subdivisions and for inter-regional correlations. In this paper, I will summarize the results of my field research and the data of numerous publications concerning the rapid changes of climate during the Pliocene–Pleistocene. The data on the Pliocene and Eopleistocene will receive special attention, as my own materials refer primarily to that time period.

In reconstructing the paleoclimates of the last 5 Ma, the sections from the lower reaches of the Kolyma and Indigirka rivers supply the most representative information in Northeast Asia. These areas were far away from the Pacific Ocean and were affected almost exclusively by the Arctic Ocean. Moreover, they are located near "the pole of cold" of the Northern Hemisphere and, therefore, may preserve indications of global climate changes in their purest form. Furthermore, these sections contain ample paleontological evidence (Fig. 1) and have been extensively studied.

In this region, the deposits of the last 5 Ma are composed almost entirely of continental facies with numerous sedimentary hiatuses. When it was impossible in certain sections to fill the hiatuses with materials from other sections of the Kolyma–Indigirka interfluve, supplementary data from more eastern areas of the northern margin of Asia were invoked.

The datings of paleoclimatic events were based on biostratigraphy and magnetochronology, applying chronology in cases where paleogeographic events were either dated independently of the climate, or fit well into the relatively detailed climate-stratigraphic Siberian scale (mainly for the last 0.7 - 0.8 Ma). Numerous radiological (radiocarbon) datings for the last 40 ka exist, but only a few thermoluminescent K-Ar datings are available for the earlier stages. The older the deposits, the more deficient their chronological justification (biostratigraphic and paleomagnetic). The deposits of 5 to 2.5 - 2.7 Ma old are represented by only 7 sections, which were studied both paleomagnetically and palynologically (Fig. 1). Some of these sections contain assemblages of diatoms, and there are individual finds of theriofauna of the 2.5 - 1.8 Ma period. Dozens of the deposit sections younger than 1.8 Ma (1.8 - 0.7 Ma) were studied using paleomagnetic methods. A still larger number of sections were characterized by theriofauna and interesting finds of foraminifers, malacofauna, etc. The middle-upper

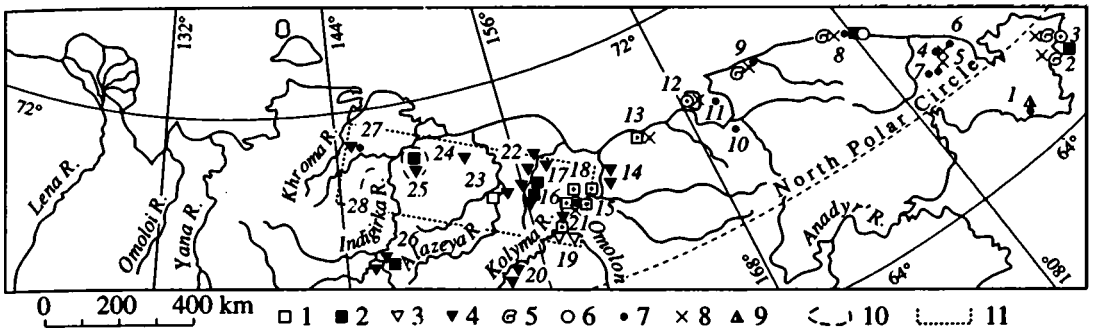

Fig. 1. A scheme of the most representative paleontological and magnetostratigraphic locations used in paleoclimatic reconstructions of the northern regions of Northeast Asia discussed in the paper.

1, 2 – Sections studied using the paleomagnetic method: 1 – Brunhes–Gauss epoch and older and 2 – Brunhes and Matuyama epochs; 3, 4 – locations of mammal fauna: 3 – Kutuyakh assemblage and 4 – Oler assemblage; 5 – marine molluscs; 6 – foraminifers; 7 – reference sections for palynostratigraphy; 8 – marine diatoms; 9 – K-Ar datings from 5.9 ± 0.4 to 4.2 ± 0.7 Ma; 10 – reference sections of the Pleistocene; 11 – the region of paleoclimatic reconstruction. Locations (indicated by numerals on the map): *1* – Enmelen River; *2* – Kukun' River; *3* – Pinakul'; *4* - *6* – holes drilled in the Vankarem basin: *4* – hole 13, *5* – hole 16, and *6* – hole 28; *7* – Pen'el'khin Creek; *8* – Reveem River; *9* – delta of the Pegtymel' R. (hole 11); *10* – Chaanai R.; *11* – Bol'shoi Routan L.; *12* – Aion L. (hole 1); *13* – Nekkeneiveem R.; *14* – area of Zelenyi Mys settlement; *15* - *18* – holes of lower Kolyma profile: *15* – hole 1, *16* – hole 2, *17* – hole 3, and *18* – hole 4; *19* – Krestovka R.; *20* – Srednekolymsk settlement; *21* – Duvannyi Yar, reference section of the Pleistocene; *22* – Bol'shaya Chukoch'ya R.; *23* – Rassokha R.; *24* – Bol'shaya Khomus-Yuryakh R.; *25* – Keremesit R.; *26* – Badyarikha R.; *27* – Tastakh L.; *28* – Allaikha R.

Pleistocene deposits contain theriofauna in great abundance, particularly in northeastern Yakutia. These occurrences are described in many paleontological and stratigraphic papers, and are therefore not shown on the map (Fig. 1). Palynologically studied sections are equally numerous, but the map presents only the important ones to be discussed in this paper.

Climatic changes were reconstructed using mostly palynological and other paleontological data. A combined study of paleontological materials with the lithofacies and paleocryological data is of considerable importance for the study of deposits younger than 3 - 2.5 Ma. As a reference point between the reconstructed climate and that of today, I would like to mention the modern climate parameters for the lower reaches of the Kolyma and for the northern part of Eastern Chukotka, respectively. For example, the average temperatures are −35°C and −25 - −26°C in January, and 11°C and 4 - 6°C in July; the average annual temperatures are −13.5°C and −12 - −14°C, and the average annual precipitation is 230 mm and 250 - 300 mm.

In the Kolyma–Indigirka interfluve area there was a hiatus in sedimentation at the Miocene–Pliocene boundary. Moreover, in this region, the accurate datings of continental facies of the latest Miocene are too scanty. An important piece of relevant information, however, was recently recovered north of the Chukotski Peninsula (Zyryanov *et al.*, 1992) from hole 16, which was drilled on the Vankarem lowland in marine deposits containing an assemblage of diatoms of the *Neodenticula kamtshatica* zone corresponding to the interval of 6.6 - 5.1 Ma (Koizumi, 1985). Zyryanov's description of the palynospectra of the pre-Hypoarctic zone is also relevant. In the northern lowlands of Northeast Asia, the Messinian crisis caused the appearance of paleotundra, which at that time, apparently, also covered Alaska, according to the oldest tundra insects recovered from clays of the Kemp lavas, and dated to approximately the same period (Hopkins *et al.*, 1971).

In Northeast Asia, the transition time from the Messinian crisis to the Pliocene optimum and the corresponding paleoclimate is poorly understood.

The Pliocene optimum coincides with the middle Beguny time, whose interval, using paleomagnetic data from the lower Kolyma profile (Fig. 1) (Paleogene..., 1989), I determined to be 5.1 to 3.4 Ma. The optimum is dated as an event 4.5 - 4 Ma old. Palynological data for this time period is both abundant and ambiguous (Volobueva *et al.*, 1990; Giterman, 1985, Paleogene..., 1989; Sher *et al.*, 1979; etc.). My results indicate (Laukhin, 1989; Paleogene..., 1989) that vegetation of the Pliocene optimum in the lower reaches of the Kolyma River was of the southern taiga type. A.F. Fradkin recognized its analogs in the region of the Great Lakes of the US (Paleogene..., 1989) and, probably, in the Southern Urals. The paleoclimate characteristics in Fig. 2 are based on these analogies. The maximal shifts of vegetation zones to the north during the last 5 Ma are reconstructed for the Pliocene optimum, when the mean annual temperatures rose above present temperatures by 10 - 12°C, and the amount of precipitation was 2 - 3 times higher. Most striking is the difference between the average temperatures in January at the optimum time and now, which in the lower reaches of the Kolyma River equals 20 - 22°C.

The palynospectra of the Chaun–Chukotka Pliocene optimum, which were recovered along with foraminifers from hole 1, located on Aion Island, were dated by T.P. Polovova back to the first half of the Pliocene (Volobueva *et al.*, 1990). The corresponding transgression may be related to the opening of the Bering Strait

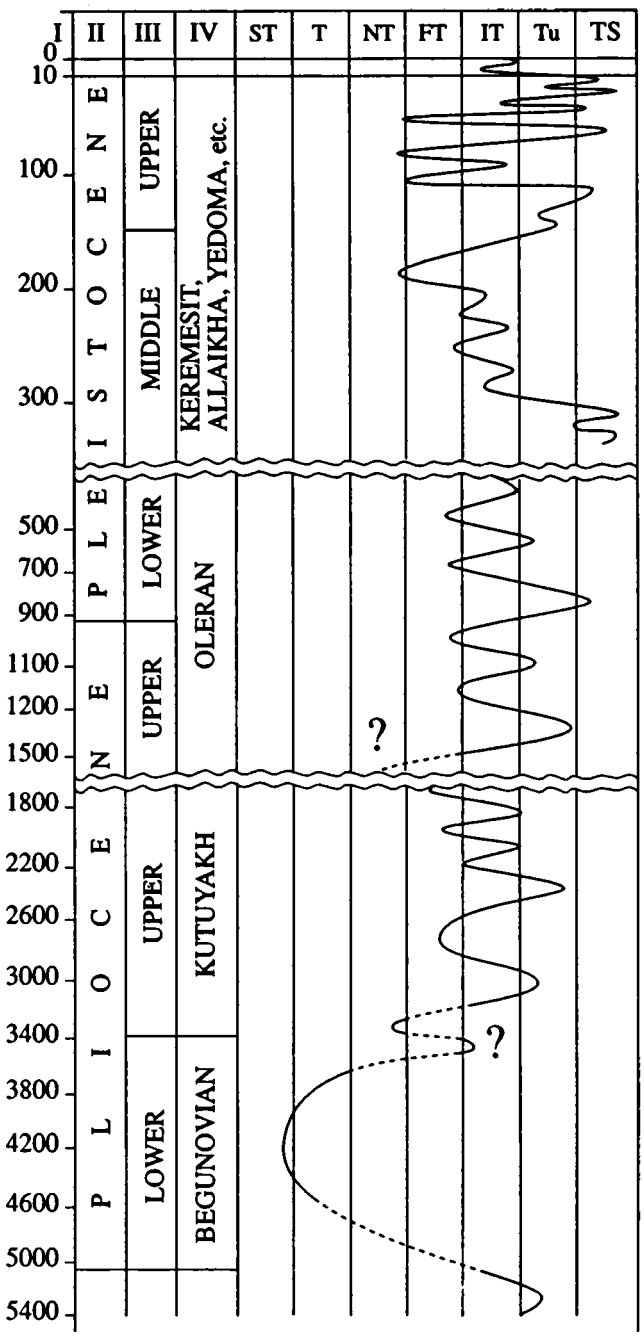

Fig. 2. Paleoclimatic curve of the last 5 Ma for the northern regions of Northeast Asia (Fig. 1). I – Time in ka; II – epoch; III – age; IV – local stratigraphic subdivisions (series, sequences, beds). Natural-landscape zones and subzones: ST – southern taiga, T – taiga, middle taiga, NT – northern taiga, FT – forest-tundra, IT – insular forests in tundra, southern tundra, Tu – tundra, and TS – tundra-steppe.

4.1 Ma ago; indications of this event were detected by Basilyan *et al.* (1991) on Karaginskii Island. The palynospectra of the Pliocene optimum were also found in the Ust'-Pegtymel' depression, together with marine diatoms similar to the Enemten and Limimtevayam assemblages of Kamchatka, along the Chaanai River, on Bol'shoi Routan Island, on Reveem River, and in other localities (Fig. 1). In the southeastern part of the Chukotski Peninsula, the palynospectra of the pine-spruce forests of the southern taiga type with *Tsuga, Abies,* and small amounts of *Corylus, Myrica,* etc., were observed in layers dated, using the K-Ar method, as 4.2 ± 0.7 and 4.3 ± 0.5 Ma old (Akinin *et al.*, 1988).

At that time, the climate became distinctively milder on nearly all the continents. A considerable warming is indicated by the composition of Brunssum floras on the Rhine (Zagwijn and Uager, 1987). The climate of the Ukraine was almost subtropical (Sirenko and Turlo, 1980). Subtropical fauna of molluscs in the Irtysh basin was described by Zykin (1986), and Hilden (1987) showed the propagation of thermophilic *Globorotalia punicticula* into the Mediterranean Sea. The temperature of the surface waters off the coasts of Antarctica was 7 - 10°C higher than it is today (Ciselski and Weaver, 1974), causing a drastic shrinking of the Antarctic ice sheet and a rise in the level of the world's oceans.

The end of the Beguny time marks the first onset of cooling. For example, in the middle reaches of the Alazeya River (Fig. 1), the upper beds of the Gilbert epoch contain the palynospectra of larch forests with localized pine and birch strips, but with vast areas of birch and alder brushwoods, perhaps evidence of a transition to forest-tundra (Kartashova *et al.*, 1987). On the Nekeveem River, again in the upper beds of the Gilbert epoch, the palynospectra are characteristic of thin northern taiga forests and forest-tundra (Volobueva *et al.*, 1990). At the end of the Beguny time, the palynological data from the lower reaches of the Kolyma River indicate an expansion of dark coniferous (spruce) taiga followed by the forest-tundra vegetation (Laukhin, 1989; Sher *et al.*, 1979). The upper beds of the Beguny Formation have ground veins of frost cleft origin and contain the remains of tundra carabuses (Sher *et al.*, 1979). In the lower reaches of the Kolyma River, this cooling stage in the sequence of paleoclimatic events can be dated, apparently, as 3.5 - 3.6 Ma old; thus correlating to the onset of the Scott Glaciation in Antarctica (Mayevskii, 1975), the Viedma Ice Age in Patagonia (Mercer and Sutter, 1982), and the glacial expansion in the Wrangell Mountains of Alaska (Armentron *et al.*, 1978). This stage of cooling may also coincide with the traces of cooling detected in the middle of bed 12 on Karaginskii Island (Basilyan *et al.*, 1991). If this speculation is true, then the peak of cooling in Fig. 2 should be lowered to 3.8 - 3.9 Ma.

The palynospectra of the northern taiga are also typical of the uppermost beds of the Beguny Formation in the lower reaches of the Kolyma River. Short-term warming at the very end of the Beguny time is evident in the palynospectra of coniferous-parvifoliate forests, which are typical of layers underlying the lower moraine drilled by hole 28 in the Vankarem lowland, located in the northern part of East Chukotka.

In the lower reaches of the Kolyma River, the Beguny Formation is covered with Kutuyakh beds,

whose upper part in the lower reaches of the Krestovka River (Fig. 1) contains the fauna *Synoptomys* (cf. *Plioctomys*) sp., *Mimomys* cf. *coelodus* Kretz., and *Cromeromys* cf. *irtyshensis* Zazh. The evolutionary level of the *Cromeromys* and *Mimomys* dentition is typical of the middle Villafranchean, whereas the progressive molar structure of *Synoptomis* allows us to date the uppermost Kutuyakh beds back to the beginning of the middle Akchaghylian (Sher *et al.*, 1979). The age of this fauna, along with the paleomagnetic and palynological data on the Krestovka River section and the lower Kolyma profile, indicate that the Kutuyakh beds belong to the time interval ranging between 3.4 and 1.8 Ma (Laukhin, 1989; Paleogene..., 1989). In the first half of the Kutuyakh period, climatic fluctuations in the lower reaches of the Kolyma River did not contrast (Giterman, 1985). Moreover, the most thoroughly studied section of the Kutuyakh beds, located in the lower reaches of the Krestovka River, is largely deformed by cryogenic processes. This circumstance hinders the application of palynological data in paleoclimate reconstruction. The climates of the first half of the Kutuyakh period were reconstructed (Fig. 2) using data from the section discovered in hole 28, located in the northern section of East Chukotka (Fig. 1).

Hole 28 penetrated two late Pliocene moraines (Laukhin and Patyk-Kara, 1991). The early glaciation coincided with the Bering transgression, and was dated as 3.2 - 3 Ma old, the later one is 2.5 - 2.4 Ma old, which corresponds to paleoclimatic data and interregional correlation. According to more recent data, however, the Bering transgression could have occurred 3.6 - 3.2 Ma, with the most intensive cooling occurring at the beginning of this stage (Basilyan *et al.*, 1991), indicating that early glaciation in Chukotka is somewhat older than 3.2 - 3 Ma.

V.A. Klimanov calculated the paleotemperature curves and the annual precipitation for these glaciation periods by applying the information-statistics method to palynological data from hole 28 (the analyses were done by B.V. Belaya). They established that the summer temperatures during the interglacial stage differed only slightly from those of the early glaciation, but that the winter temperatures were 8 to 9°C lower (Fig. 2). Nearly identical temperatures and amounts of precipitation characterize present regions of the Saint Elias Mountains in Alaska, where the glaciers reach down to sea level. In northern Chukotka, the present climate is more severe, but glaciation cannot develop because the precipitation is less than half of that in the Saint Elias Mountains. It seems obvious that the early glaciation was initiated not so much by the temperature decrease as by an increase of humidity, i.e., as a result of the marine (Bering) transgression and heavier precipitation. In the central regions of East Chukotka, far from the sea, the most ancient pseudomorphs along the ice veins on the Pen'el'khin Creek may correspond to the early glaciation (Laukhin *et al.*, 1988).

Traces of glaciation from about 3.2 Ma are found in Iceland (Akhmet'ev *et al.*, 1978), in the Pamirs (Nikonov and Pakhomov, 1984), in Patagonia (Mercer and Sutter, 1982), in the Cordilleras (Curry, 1979), and in other regions. About 3.6 Ma, the glaciers formed on the Wrangell Range (Armentron *et al.*, 1978). In Antarctica, the Scott Glaciation began to form about 3.5 Ma (Mayevskii, 1975). This cooling caused the appearance of *Globorotalia crassiformis* (Drawry, 1979) in the Mediterranean, and the first paleoarctic molluscs in Southern-western Siberia (the Irtysh River), (Zykin, 1986). The more recent glaciation (2.5 to 2.4 Ma) in northeast Chukotka occurred when winter temperatures were 16 - 20°C lower than present temperatures in the region of mountain–valley glaciation in Alaska (about 60°N), where the annual precipitation is presently 50 - 100 mm or greater. It seems obvious that the more recent glaciation in the Pliocene in Northeast Chukotka originated mainly as a result of the temperature decrease. In the lower reaches of the Kolyma River, this time period correlates, it seems, with the formation of polygonal vein ice in one of the pseudomorph systems of the Kutuyakh beds. Glaciation in the Alps (Carraro *et al.*, 1975), Caucasus, and Sierra Nevada Range (Curry, 1979), along with the Elk Creek Glaciation in Nevada, Nebraska, and Iowa (Nikiforova, 1989) occurred in approximately the same time period. Nikiforova (1989) dates the oldest loess in the South Pamirs, the Tashkent region, and China to nearly the same period. In this same period the first lemmings appeared in Europe (Azzaroli, 1983), in North America (Reppening, 1989), in Eastern Siberia (Agadzhanyan and Erbaeva, 1983), and in Western Siberia, also where the first pseudomorphs formed in the ice veins (Zazhigin, 1980).

In the third and final stage of the Kutuyakh period, the climate in the lower reaches of the Kolyma River oscillated; the forest-tundra vegetation repeatedly changed to tundra flora, and polygonal-vien ice appeared and then degraded. Individual stages of warming and cooling during that time were palynologically studied at various sections between the Kolyma River and the Bering Strait (Fig. 1), but it is almost impossible to fix their exact time within the interval from 2.3 to 1.8 Ma. On the whole, this stage may correspond to Europe's Tegelen, which also shows a series of warmings and coolings.

In the lower reaches of the Kolyma River, the interval from 1.8 to 0.6 Ma corresponds to the Oler time. Between the Kolyma and the Indigirka rivers, the Oler Formation, which is fairly well studied paleomagnetically, has numerous occurrences of mammals (Sher *et al.*, 1979; Laukhin *et al.*, 1983; Resheniya..., 1987, etc.). There are two assemblages of the Oler fauna, namely, the lower one (1.8 - 0.8 Ma) with *Predicrostonix compitalus* Zazh., *Allophaiomys* cf. *pliocenicus* Korm., *Gulo* cf. *schlosseri* Krom., *Arctelephas* sp. 1, *Praeovibos beringiensis* Sher., *Homoterium* sp., etc.; and the upper assemblage (younger than 0.8 Ma) with *Dicrostonix renides* Zazh., *Arctelephas* sp. 2, *Equus* ex gr. *mosbachensis* Reich., *Cervalces* af. *latifrons* Johne., *Soergilia* af. *elisabethae* Schoub., *Canis lupus* cf. *mosbachensis* Kretz., *Praeovibos* cf. *priscus* Staud., *Rangifer* sp., etc. As indicated by paleomagnetic, pale-

ocryological, and palynological data, the Oler time can be divided into three intervals: early, middle, and late.

In the beginning of Oler time, there was a warming, evident from palynological data, at the boundary with the Kutuyakh event. This warming corresponds to the deposition of peat bogs at the base of the Oler Formation in the Bol'shaya Chukoch'ya river valley. It occurred nearly at the same time as the warming at the end of Olduvai epoch. The warming resulted in forest propogation far north of Greenland during 1.8 - 1.6 Ma (Funder *et al.*, 1985), and in the positive winter temperatures that occurred for the last time during the Cenozoic in the northern Ukraine (Sirenko and Turlo, 1980). In the early Oler time, an intensive cooling preceded warming, and in the lower reaches of the Kolyma River, the first tundra assemblages appeared at this stage of the Cenozoic. This succession of paleoclimatic events during the early Oler period correlate with the Eburon period in Europe, and with the Middletonian in North America.

In the middle Oler time, about 1.3 - 0.8 Ma, peat bogs first appeared in the larch-birch open woodland, and the permafrost partially degraded. The pseudomorphic ice veins indicate a low cooling peak. Another wave of cooling marks the end of the middle Oler interval, resulting in the development of tundra-steppes in the lower reaches of the Kolyma River. The coeval regional horizon containing Chenopodiaceae lies below the Brunhes–Matuyama boundary, overlapping it in some sections (Resheniya..., 1987). During that period of cooling in Northern Chukotka, the early Enmakai transgression began to develop into the next warming event. In the Val'karai basin of the Reveem River, the lower Enmakai beds are dated as 0.93 - 0.63 Ma old according to paleomagnetic data and assemblages of diatoms and foraminifers (Fig. 1). The cooling that corresponds to the Chenopodiaceae horizon approximates the Gunz (Menapian) and the Nebraskan events in North America in time and intensity. In Europe at that time, the first ice sheet was formed, and the Scandinavian ice cover, according to Zubakov(1990), was as large as the Valdai ice sheet. Volkova (1981) reconstructed the position of the Western Siberian vegetation zones of that time to show that, compared to their present location, they were shifted 9° - 10° southward.

There were two coolings in the late Oleran (0.8 - 0.6 Ma), both above the Brunhes–Matuyama boundary, as testified by paleocryological data from the upper Oler member in the Kolyma–Indigirka interfluve. The palynological evidence indicates that the larch-birch forests and forest-tundra vegetation of that time changed later to tundra. The exact age of these events is unknown, and it is unclear whether there was one "warming–cooling" cycle or two, as shown in Fig. 2. The two cycle theory is based on paleocryological data. The palynospectra that reflect coolings have different stratigraphic positions in the upper Oler beds in the Kolyma–Indigirka interfluve. Moreover, there is data on the asynchronous Avlekat (580 ± 150 ka) and Mavrin (470 ± 120 ka) glaciations in the Urak plateau in the North Okhotsk region, obtained by the thermoluminescent method (Resheniya..., 1987). Traces of two glaciations (0.8 - 0.45 ka) can be detected in Kamchatka and in the upper reaches of the Kolyma River, though they are dated by a less reliable paleomagnetic method. This time period corresponds to the Gunz–Mindel stage, which correlates with the second half of the Enmakai transgression found in Chukotka, and then to the Mindel glaciation in Europe, and to the Kansas glaciation in North America. The above events correspond to the Mansi and Shaitan ice ages in Western Siberia.

The middle Pleistocene deposits and the paleoclimates of that epoch are best studied in the lower reaches of the Indigirka River along the Keremesit and Allaikha river valleys (Fig. 1). In these and many other sections, the middle Pleistocene beds contain an abundant fauna of mammals: *Ochotona hyperborea* Pall., *Equus* sp., *Mammuthus* sp., *Bison priscus* Boj., and other small and large forms.

At the beginning of the middle Pleistocene, the northern birch-larch taiga forests moved north, practically reaching the Indigirka delta. In this period, permafrost degraded for the last time, perhaps entirely, to the south of the present coast line. During the optimum, the mean annual temperatures may have risen to 0°C. This stage corresponds to the Mindel–Riss and Likhvin Interglacials in Europe, and the Yarmouthian Interglacial in North America.

Data from the Kolyma–Indigirka interfluve clearly indicates two coolings in the middle and in the second half of the middle Pleistocene. Both events resulted in the appearance of cryoarid tundra-steppes. They are characterized by numerous palynospectra, and cryoxerophyte entomofauna characterized by *Dicrostonix simplicior* Fejf., *Lemmus* cf. *obiensis* Br., etc. The beds with *Dicrostonix* in the Achchagai–Allaikha River were deposited at that time. Using the oxygen–isotope composition ($\delta^{18}O$) of the polygonal ice veins, Arkhangelov *et al.* (1990) determined that average winter temperatures were −37 - −40°C in the last middle Pleistocene cooling. Palynological data and several levels of cryoturbation indicate that the cooling period was complex and had more stages than shown in Fig. 2. There were at least two glaciations in the mountains of Chukotka. During the warming in between, the mean annual temperatures may have been by 2 - 5°C higher than today. At the same time, the forest-tundra and northern taiga vegetation moved northward. The permafrost partly degraded, and peat bogs appeared. The climate during that period was interglacial, which the glaciers in the mountains of Chukotka just barely survived. This stage corresponds to the Riss event in Europe, and correlates with both the Samarovo and Taz glaciations in Siberia, and with the Illinois Glaciation in North America.

A succession of late Pleistocene paleoclimates was studied using numerous exposures of the ice-derived (Yedoma) deposits of the Kolyma–Indigirka interfluve. In the lower reaches of the Kolyma River, 0.095 - 0.12 Ma, the larch-birch forests appeared again,

and prograded up to 300 km north of the present northern boundary of forest vegetation. The southern boundary of permafrost retreated almost to the present-day coast line; the average temperatures in January and July were 2 - 4°C higher than at present, and the precipitation was 100 - 200 mm higher than present levels (Velichko and Nechaev, 1988). This stage corresponds to the Riss–Wurm and Mikulino Interglacials in Europe, the Kazantsevo Interglacial in Siberia, and the Sangamonian Interglacial in North America.

In the Kolyma–Indigirka interfluve area, three stages of sharp cooling during the period 0.01 to 0.095 Ma were recognized, namely during 95 - 50, 34 - 29, and 24 - 10 ka. The mean annual temperatures probably fell to –25°C; after Arkhangelov *et al.* (1990) the mean winter temperatures were –40 - –43°C whereas during 24 - 10 ka they were –39 - –42°C. Based on paleocryological data, Kaplina and Kuznetsov (1975) have estimated that the mean annual temperatures of the late Pleistocene coolings were between –20 - –32°C, thus implying that the mean temperatures in January were –70 - –80°C. The actual values are, apparently, somewhere in-between, but probably not above the mean annual values presently found in Oimyakon (about –17°C). The glaciation in the mountains of Chukotka occurred in several stages. The further the glaciation's eastward progress, the closer it extended towards the foothills. The peat bogs, fossil soils, and thermokarst developed during the warm periods between the coolings which occurred 34 - 50 and 24 - 29 ka ago. The arborescent vegetation reached the ocean 34 - 50 ka ago, and in the Kolyma–Indigirka interfluve the mean temperature for January was –32°C and for July +15°C; the annual rainfall was as much as 250 - 300 mm. Presently, many radiocarbon datings of this stage synchronize it with the Wurm in Europe, the Wisconsinian in North America, and the Zyriansk and Sartan Glaciations in Siberia.

To conclude, during the last 5 Ma, the fluctuations of climate in the far north of Northeast Asia were numerous and of large amplitude, particularly in the Pleistocene. The sequence of climatic events in general closely corresponds to climatic events in the Northern Hemisphere, and has analogs in the Southern Hemisphere. Knowledge and in particular the datings of climatic events in Northeast Asia, however, are still far from adequate.

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Geochronology of Weathering Crusts on Flood Basalts in Syria, and the Evolution of Regional Paleoclimate during the Last 20 Ma

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Abstract – A method and results of weathering crusts (WC) dating, were studied. The dated WC are considered indicators and time reference points of definite climate conditions. This concept prompted our study on inner (intralava) WC in flood basalts of various ages in Syria. It is established that a montmorillonite eluvium formed in the region 21 Ma to 9 Ma; a gibbsite–halloysite–goethite eluvium is dated as 5.5 - 4.5 Ma old. The formation of the WC ceased at the Pliocene–Quaternary boundary. The evolution of the WC indicates significant climate fluctuations in the region from semiarid (early, middle, and late Miocene) to humid warm (early Pliocene), and, finally, to dry arid conditions (Quaternary). The isotopic-geochronological data indicates that the formation of the WC is, geologically, an instant process (time intervals from a few hundred ka to 1 Ma).

As is known, weathering crusts (WC) are only formed under terms of hypergenetic processes. Hence, they are paleoclimate indicators of the Earth's geological history. Climate plays a definite role in the formation of various WC mineral compositions. Which is to say, depending on the landscape and climate, the same rock is capable of producing different weathering products. Regional WC, formed simultaneously in various climate belts, has a zonation expressed by the regular distribution of its mineral types in space (Strakhov, 1962). In turn, the regionally forming thick eluvial cover changes various relief features of the Earth's surface (peneplanation, changes in vegetation cover, etc.) and leads to significant modifications in the environment and climate. Over time, climate changes are reflected by changes in WC mineral types, or its absences along a stratigraphic sequence. Thus, data on distribution regularities, age, and WC mineral composition may be used as indicators and time reference points of both the lateral and vertical climatic and physical-geographical fluctuations recorded in the studied sections.

Until recently, the "qualitative" time intervals, corresponding either to eras or more frequently to periods and epochs and rarely to stages, were an accepted means of defining WC age. Similar datings (Mesozoic, Cretaceous, late Miocene WC etc.) were too uncertain to understand the actual period of WC formation (from hundreds and to more than several Ma), and their precise geochronology. Numerical data on WC ages (Novikov *et al.*, 1989, 1991) began to appear in publications only recently. We conducted a geochronological study of WC in the late Cenozoic flood basalts of Syria.

METHODS

WC within areas of recent volcanism are ideal objects for age determination. The cyclic character of flood basalt formation resulted in a development of inner (intrabasalt) WC during pauses between magmatic activity, and then in their conservation after the magmatic regeneration. As a result, the multilevel layered sequence, consisting of alternating flows of unaltered basalts and the rocks replaced by clayey eluvium, was formed. The WC age can be estimated using K-Ar datings on the host rocks. The early and late WC age limits were established by sampling basalts at the base and above the fossil eluvium bed. The difference between both ages defines the interval of WC formation. To measure ultralow contents of radiogenic Ar (about 10^{-10} - 10^{-11} g) we used a complex apparatus of a low-background type: MI-1330 mass spectrometer working in a static regime with an automatic system of operation control and registration, based on PC "Elektronika DZ-28," and a modernized A-4-7 separation unit. The background concentration ("hot" blank exper-

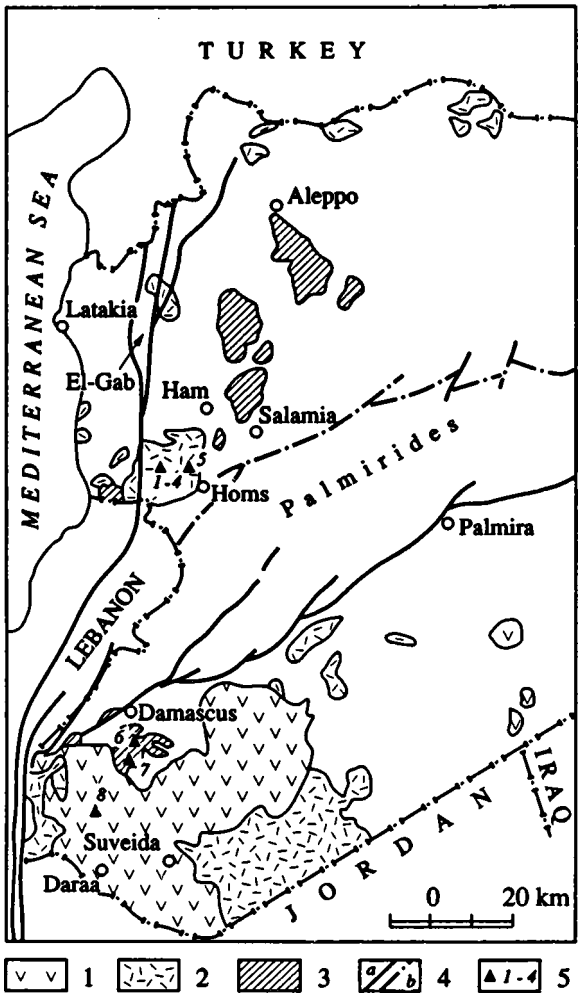

Fig. 1. Scheme of flood basalt distribution in Syria and localities of sampling for isotopic dating. 1 - 3 – basalts: 1 – Quaternary, 2 – Pliocene, and 3 – Miocene; 4 – established (*a*) and supposed (*b*) faults; 5 – location and numbers of studied sections.

iment) was (6 - 8) × 10^{-12} g on ^{40}Ar, being, as a rule, two orders in magnitude lower than amount of radiogene Ar in the experiment. The content of radiogenic Ar was measured using isotopic dilution with an ^{38}Ar tracer; the measurement precision was controlled by periodical measurements of standard samples and the Ar isotope composition in atmosphere. The potassium content was measured using flame-photometry.

The study was conducted on the Pliocene Shin Plateau (Homs Area), and in outliers of Miocene basalts of the Hauran–Druza Lava Plateau south of Damascus (Fig. 1). Interbasalt WC are present in both flood basalt regions. This provides us with a unique opportunity to date the WC of various ages formed within the limits of one region during a significant time interval (about 20 Ma), and to trace the evolution of these rocks, in order to reveal the paleoclimatic conditions at the time of their formation. It is important to point out that WC occur neither on the surface, nor within flows of Quaternary basalts distributed widely throughout this region.

GEOLOGICAL FEATURES OF WEATHERING CRUSTS IN THE STUDIED BASALT PLATEAUX

The northwestern part of the Arabian plate is a classic region of flood basalts, which appeared during eruptions occurring in the period beginning 20 Ma ago and ending in the historical period. Their origin is related to geological activity in the northeastern periphery of the Red Sea Rift Province. The lava plateaus are formed by high-Ti basalts of moderate and normal alkalinity. The scale of lava fields varies widely from a single valley-flow to vast lava plateaus that occupy areas up to hundreds, thousands, and even hundreds of thousands of square kilometers (Syria–Jordan Plateau). The thickness of such lava sequences may reach 1 - 1.5 km, and the number of flows, several dozen. The chains of scoria and scoria-pyroclastic cones, tracing fractures generally orientated northwestward and used as feeding canals, are frequently observed on the surface of lava plateaus.

The late Cenozoic magmatism in this region developed in three stages: (1) the early Miocene stage corresponding to the opening of the Red Sea Rift; (2) late Miocene–early Pliocene stage, contemporaneous with the opening of the Akaba Bay, and the formation of the Dead Sea (Levant) Transform Fault; and (3) the late Pliocene–Quaternary stage, characterized by a general uplift of the territory, and the modern relief formation. The distribution of basalts of various ages on the territory of western Syria is shown on Fig. 1. We investigated samples of all three age groups.

WC within the Pliocene Shin Plateau have been comprehensively studied. The plateau is adjacent to the Levant Fault, clearly visible in northwest Syria as the El-Gab Depression (Ponikarov *et al.*, 1969), and represents a northwestward elongated upland. It has an oval shape in plan and occupies about 1200 km^2. According to borehole data, the relief of the plateau base is uneven. The thickness of the basalt cover in the central part is about 350 m, and decreases to as little as 50 m towards the margins. The plateau is formed by a series of basalt flows, up to 25 - 30 in number in the plateau central part. Earlier, it was supposed that subhorizontal lava flows could be traced throughout the plateau area. However, a preliminary study of the space images for Shin Plateau, the Homs area, revealed a presence of at least several eruptive centers elongated along the plateau axis. Apparently, they did not act simultaneously, because the general stratigraphic sequence reconstructed on the basis of local sections (Sharkov *et al.*, 1987) is inconsistent with the results of our detailed geochronological study. The thickness of lava flows varies from 0.5 to 10 - 12 m. They alternate with WC in the sequence (Sharkov *et al.*, 1986).

WC in the Homs area occur widely. They are seen in natural outcrops, in road cuts, and were also recovered by boreholes, represented as subhorizontal aprons located mainly between bazalt flows. As a rule, on the surface on the uppermost basalt flows they are absent. Thickness of the eluvial formations does not usually exceed 0.5 - 1 m, although in some cases it is as much as 45 m. Geochronological dating of WC was conducted in the central part and eastern margin of the Homs Basalt Field. Four reference sections were selected in the central part of the Shin Plateau: approximately 20 km to the west-northwest of Homs City near the Bahur settlement; 5 km to the northwest from the latter (In-Alkud Mountain); and 1.5 and 4 km eastward (the Ein-Zagra and Hadie settlements, respectively). The first section represents a thick, about 100 m, rock-unit, in which there are 10 thin WC layers (each 1 to 8 m thick). The host basalts of the thickest upper eluvial horizon (samples 66-11, 66-12), and basalts of the WC base from the lower part of the section (sample 66-1) were selected for isotopic analysis. The sampling allowed us to estimate the ages of the two individual WC horizons, and to determine the age interval of the 100 m thick layered basalt-eluvial unit as a whole. Basalts, enclosing a thin (2 m) WC layer (samples 128-2, 128-1), were analyzed in the second section. The third section was selected because the WC is exposed at the day surface, a rare case in the studied region. This 2 m thick eluvial cover is supposed to be one of the last in its formation time. Basalts of the eluvium base, recording the initial stage of weathering (sample 77-4), were selected for analysis. The fourth section is distinguished by its unique WC thickness, which is 45 m, although the presence of tuffaceous material should not be excluded, as it covers the weathered basalts of the section (samples 148-3, 148-1). WC were also studied on the eastern flank of the Homs area, 20 km SSW of Hama near the Al-Zaraa Settlement. Here, under the Quaternary deposits 3 - 5 m thick several shallow boreholes recovered basalts with a layer of clayey eluvium. We studied the WC section 2 m thick. The overlying basalt flow is about 20 m thick, and the underlying flow is 5 m thick (samples 813-3, 813-1). As the section under consideration is near the base of the plateau, we assume that underlying basalts and related WC have an older age. The studied sections are shown on the Fig. 2.

WC of these and other sections of the Shin Plateau form subhorizontal interbasalt layers and consist of a homogeneous, soft, reddish-brown, clay-ocher substance. The material is plastic in a wet state, or dense and heavy when dry. As a rule, the relic structural-textural features of parent rocks, the basalts and tuff-breccias with tuff-agglomerates in single cases (section 4), are preserved. Visible zoning of the weathering profile was not observed in the studied crusts. A disintegration of basalts into separate blocks accompanied by the development of clay mineralization along the jointing net is typical of the lower WC parts. Relic blocks of the former show spheroidal jointing and become loose. Disintegrated basalts have a grey or grey-yellow color.

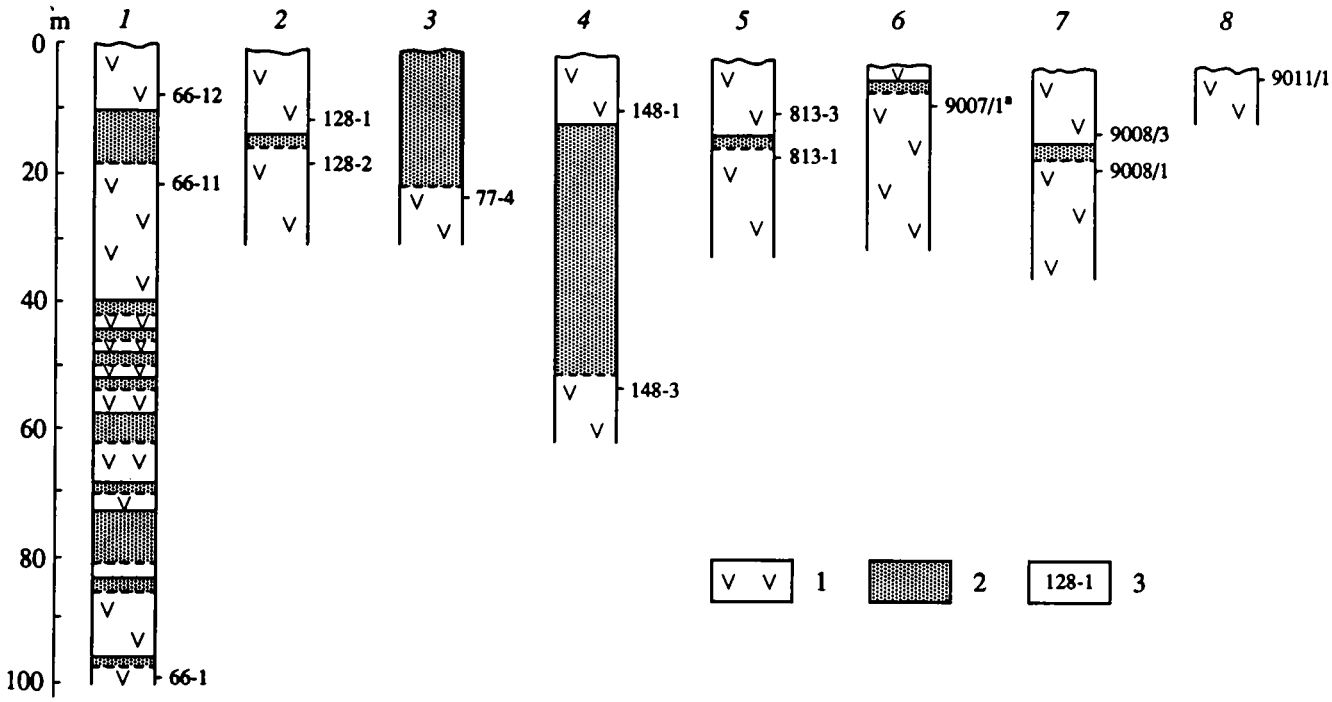

Fig. 2. Sections of layered basalt-eluvial units (numbers of sections are shown on Fig. 1). 1 – Basalts; 2 – weathering crusts; 3 – sample numbers for K-Ar dating, corresponding to those in the table.

Upwards, they grade into red-yellow and red massive lateritelike rocks.

The mineral compositions of WC in the central part of the plateau and its periphery are different. In the first case, halloysite, goethite, and hematite compose the WC. Maghemite, psilomelane, allophane, and gibbsite develop as an admixture (Sharkov *et al.*, 1986). There is a definite regularity in the distribution of hypergene minerals. The lower part of the WC consists predominantly of halloysite. Minerals of the Fe-hydrooxide group and, to a lesser degree, the alumina minerals prevail in the composition of the upper part. The chemical composition of weathering products also varies along the profile. From the base upwards, there occurs: the oxidation of ferrous into ferric oxides, the loss of the majority of rock-forming components, and an accumulation of Fe_2O_3 and Al_2O_3 in the eluvium top part. The WC of the eastern flanks of the Shin Plateau clearly express a monozonal profile type, and consist mainly of montmorillonite. Goethite, calcite, and quartz are observed in small amounts. An insignificant enrichment in hydrous ferric oxides is fixed in the upper part of the profile, where they are dispersed as a coloring agent in the clayey mass. Published data on the presence of laterite bauxites in WC of basalts in Syria exists. However, this data has not been confirmed factually (Alias Kamal Assad, 1985). So, the WC of the central part of the Shin Plateau should be classified compositionally as allite or laterite, whereas eluvium of the plateau flanks may be attributed to siallites.

The early Miocene basalts and associated WC have been studied in the Hauran–Druza Basalt Plateau. Located in southwestern Syria, it continues onto Jordanian territory. The early Miocene basalts of the plateau lie over the Cretaceous and Paleogene deposits unconformably, and, in turn, are overlain by Pliocene–Quaternary basalts, and Quaternary continental deposits of the Damascus Basin. The area, covered by the early Miocene basalts at the present-day erosional surface, is about 200 km^2. But the former size of the plateau was significantly larger. The basalt sequence under consideration consists of 20 - 25 flows separated often by WC horizons, which are widely distributed in the area. The prevailing varieties of basalts have normal and moderate alkalinity. The plateau thickness is as much as 400 m. A geochronological study of basalts and associated fossil eluvium was carried out at the northern margin of the plateau. Volcanic rocks enclosing the uppermost (latest) WC beds were selected for isotopic study. This eluvial cover is partly exposed at the day surface of plateau, but, in general, buried under thick basalt flows. The aim of the dating was to estimate the age of the WC horizon, marking the time of terminal Miocene volcanism. Two similar sections situated at similar hypsometric levels 10 and 30 km to the south of Damascus (see Fig. 2) were selected for our study. The first section studied in the escarp near the Jabal-al-Kalb road consists of the following members, from the top downwards:

(1) Loose weathered basalts with calcite, zeolites, and carbonate geodes in hollows; in the upper part of the member there are horizontally distributed carbonate concretions; 2 - 3 m thick.

(2) WC in layers and lenses of irregular shape; 1 m thick.

(3) Black, hard basalts; 3 - 4 m thick.

The second section is situated near the village of Menkat al-Hatab. Two basalt flows with an eluvial horizon between them are exposed in the road quarry. The WC and host rocks (upper and lower) are as thick as 2, 3, and over 5 m, respectively.

In the first section, we analyzed basalts of the lower flow (sample 9007/1a) and in the second section, rocks of the upper and lower flows (samples 9008/3, 9008/1).

The WC of the Hauran–Druza Basalt Plateau are similar to the eluvium of the Shin Plateau in their mode of occurrence and external appearance. They also form subhorizontal aprons within the basalt sequence. The lower boundary (base) of the WC containing maternal rocks is uneven and sharp. The clayey eluvium forms a horizon of a consistent thickness at the top of the

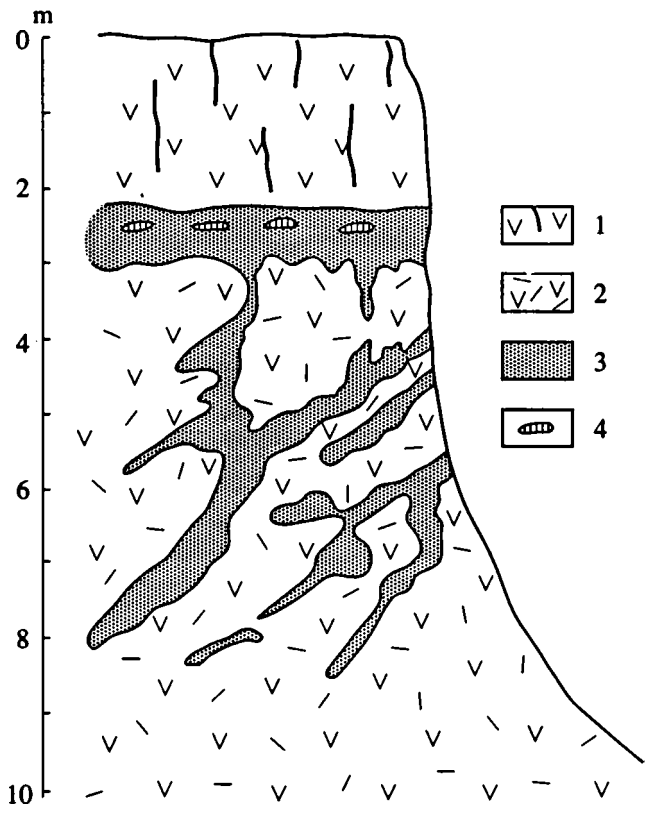

Fig. 3. Sketch section of an interbasalt eluvial horizon, characterizing interrelations with the host rocks. 1 – Unweathered and 2 – fractured weathered basalts; 3 – weathering crusts; 4 – carbonate concretions.

WC-bearing flow; in its lower part, it develops along jointings as pockets and "veins" penetrating down to 6 - 7 m. The upper contact of the WC with overlying basalts is distinct and flat (Fig. 3). The weathering products are represented by fine-grained and homogeneous red clay. The clay consists of montmorillonite, calcite, quartz, and goethite. There are often carbonates concretions.

Our studies show that the WC in Syria, attributes of the Pliocene and Miocene basalts, are unknown among or above the Quaternary volcanics. This is evident, first of all, in various geomorphologies of basalts fields of different ages. The Quaternary and, especially, the Holocene volcanics are represented by well-preserved cinder volcanoes and sheets with a typical volcanic relief on the surface. The Pliocene and Miocene effusives do not reveal a clear, originally volcanic morphology; their relic eruptive centers are preserved as low-hill outliers, and the sheets are intensively denudated along their margins, revealing the WC remnants. Accordingly, the Quaternary basalts were selected for radiometric study in order to estimate the upper age limit of the clayey eluvium formation. The rocks were sampled in the young plateau, 42 km southwest of Damascus (sample 9011/1). The age the WC occurring at the day surface is older than that of the sampled Quaternary effusive rocks.

Results of the K-Ar analysis of Quaternary, Pliocene, and Miocene basalts and conclusions concerning the estimated age of associated WC are shown in the table.

DISCUSSION OF RESULTS

A consideration of the available data permits us to suggest some conclusions on the geochronology, distribution regularities, and original environments of the WC in Syria.

WC form horizontal argillaceous layers of consistent thickness between basalt flows, which conform to the bedding. The zonal structure of eluvium, gradual transitions from underlying basalts into products of their hypergene alteration, and sharp boundaries of the latter with overlying basalts are established. The character of the studied weathering profiles, compositional and geochemical peculiarities of rock-forming components, and relic pseudomorphous textures testify

Geochronology of weathering crusts

Number	Number of sample	Location	Contents in sample		$^{40}Ar_p$ in experiment, %	Age of basalts, (±1 δ), Ma	Age of weathering crusts, Ma	
			K, %	$^{40}Ar_p$, ng/g				
1	9011/1	42 km SW of Damascus	1.48 ± 0.03	0.031 ± 0.015	1.8	0.3 ± 0.2	WC not formed	
2	128-1	Ein-Kut	1.12 ± 0.03	0.371 ± 0.008	44.6	4.76 ± 0.17	late Miocene–early Pliocene	4.76 - 5.23(0.53)
3	128-2	"	1.23 ± 0.03	0.446 ± 0.007	55.7	5.23 ± 0.15		
4	148-1	Ein-Zagra	0.91 ± 0.02	0.307 ± 0.006	45.3	4.85 ± 0.15		4.85 - 5.64(0.79)
5	148-3	"	0.81 ± 0.02	0.315 ± 0.006	46.6	5.64 ± 0.18		
6	66-12	Hosur	0.85 ± 0.02	0.299 ± 0.006	37.3	5.07 ± 0.16		5.07 - 5.19(0.12)
7	66-11	"	1.18 ± 0.03	0.426 ± 0.009	47.3	5.19 ± 0.17		
8	66-1	"	0.79 ± 0.02	0.339 ± 0.007	44.8	6.2 ± 0.2		6.2(6.2 < WC < 5.19)
9	77-4	Haddie	1.06 ± 0.03	0.394 ± 0.012	26.3	5.4 ± 0.2		5.4(<5.4)
10	813-3	20 km SSW of Ham	0.55 ± 0.02	0.242 ± 0.013	12.8	6.3 ± 0.4		
11	813-1	"	0.50 ± 0.02	0.338 ± 0.014	15.9	9.7 ± 0.6		6.3 - 9.7(3.4)
12	9007/1a	10 km S of Damascus	0.54 ± 0.02	0.761 ± 0.011	55.6	20.4 ± 0.8	early	20.4 (<20.4)
13	9008/3	Men Kat	1.55 ± 0.03	2.20 ± 0.03	88.3	20.4 ± 0.5		
14	9008/1	"	1.46 ± 0.03	2.13 ± 0.03	80.3	20.9 ± 0.5		20.4 - 20.9(0.5)

unquestionably to the eluvial, as opposed to the sedimentary or hydrothermal origin of these formations.

Two age intervals of the WC formations are established: the late Miocene–early Pliocene and the early Miocene. They correspond to 9.7 - 4.76 Ma and 20.9 - 20.4 Ma, respectively. One can not exclude the possibility that the discovery of these geochronological intervals of WC formation was predetermined by the presence of basalt flows of corresponding ages.

As one would expect, the age of the eluvial horizons in the Shin Plateau is not consistent throughout its various hypsometric levels (top, foot). The formation of the WC was defined for 5 million years by the cyclic character of the basalt eruptions in the Homs area. The interbasaltic WC mark a cessation of the intervals of magmatic activity. Two such intervals are 9.7 - 6.3 Ma in the basal part of the Shin Plateau sequence, and 5 Ma, approximately, in its upper part. The latter interval is recorded in four sections of the studied area. This interval, in turn, is sometimes divided into series of more short-term pauses marked, as a rule, by thin eluvial horizons (see section 1, Fig. 2). The upper age boundary is 4.67 Ma for the Pliocene interbasalt WC.

The lower age limit of the early Miocene WC formations, studied in two separate sections of the Hauran–Druza Plateau, is essentially similar. This fact suggests a similarity between selected sections. A "stability" of the lower age limit of the early Miocene WC is apparently caused by a cessation of the basalt volcanism at that time, approximately 21 Ma. The WC was formed on the surface of effusive rocks, marking the age limit at approximately 21 Ma. The overlying basalt flows reveal a slightly younger age than the studied sections. These flows are thin, and apparently were the last ones erupted at this stage. It is significant that similar ages for the upper flows of the same basalt area were reported by Giannerini *et al.* (1988). Judging by the presence of eluvial horizons in other parts of the sequence, there were several breaks in the eruptions.

Age intervals of the interbasalt WC formation, corresponding to the breaks of volcanic activity, varied from 0.12 to 3.4 Ma. Laterite beds up to 0.5 - 45 m thick developed during these time intervals. It is unclear whether the thickness of the WC is dependant on its age. For example, the formation times of clayey eluvial layers 2, 8, and 45 m thick in the central part of the Shin Plateau are 0.53, 0.13, and 0.79 Ma, respectively. The age interval of the WC horizon inside the lower flows of the plateau is estimated as 3.4 Ma, and an eluvium of similar thickness in the Miocene basalts of the Hauran–Druza Plateau was formed over a period of 0.5 Ma, approximately.

The studied WC developed after similar rocks (basalts) in all the considered cases, excluding section 4 (Fig. 2). Their formation was simultaneous, and took place under similar geomorphological and paleoclimatic conditions in the central part of the Shin Plateau. In consequence, we conclude that the upper age limit of the WC is apparently older than the age of the overlying basalts, i.e., the WC were formed in barely a moment, in a few thousand years – the "seconds" of geological time. Moreover, it is likely that when basalts were subjected to weathering immediately after their eruptions on the Earth's surface. If the processes did not occur immediately, the time of the WC formation was shorter. Physical-mechanical properties (porosity, unhomogeneity etc.) of tuffaceous material, terminating the basalt flow in the section 4, may explain the significant difference in WC thicknesses between the second and fourth sections (2 m and 45 m) (Fig. 2), but nearly equal time intervals of their development (0.53 Ma and 0.79 Ma respectively). Its weathering should occur, apparently, much quicker in comparison to homogeneous massive basalts of other studied profiles. As mentioned above, ten eluvial horizons were recognized in the first section of the Shin Plateau. The total thickness of the layered basalt-eluvial unit is about 100 m. The estimated time interval of its formation is only a million years.

Data on the mineral composition of the WC formed after the Syrian basalts shows that it changed with time. The early Miocene (20 Ma) and late Miocene (6.3 - 9.7 Ma) WC are mainly of montmorillonite composition. The early Pliocene WC consist mainly of goethite and halloysite. Gibbsite is observed as an admixture (Sharkov *et al.*, 1986). Such a sharp change in the WC mineral composition took place, it seems, at the Miocene–Pliocene boundary (about 5 - 6 Ma), and was caused by a change in the conditions of their formation. As is known, peculiarities in the formation of montmorillonite (smallite) and gibbsite–halloysite–goethite (laterite) weathering profiles are different. The first type is developed under conditions of slow water exchange in a warm and relatively dry climate; the second type is typical of humid subtropic and tropic climate environments (Nikitina *et al.*, 1971; Sinitsyn, 1976; Strakhov, 1962). In this context, it is important to note that significant geological events, marking the transition from the Messinian transgression to the Pliocene regression, took place in the Eastern Mediterranean at the late Miocene–early Pliocene boundary. Clearly, the formation of the Messinian clay-gypsum-bearing beds, which cropped out in the coastal area near Latakia, took place under relatively arid conditions and at low sea level. As a result of the Pliocene transgression, the marine basin in northwest Syria prograded far into the land along the Al-Kebir depression. A wide freshwater basin existed in the El-Gab Rift Valley during the Pliocene and was probably connected with the marine basin in the northern part of the depression. According to palynological data (analyst A.N. Simakova) on the marine Pliocene sections of the Al-Kebir depression, prevaling coniferous pollen and the presence of such exotic forms as *Tsuga, Podocarpus, Liquidambar, Carya,* and *Magnolia* are typical of spore-pollen spectra that indicate humid and relatively warm climates.

An absence of WC, which developed after the Quaternary flood basalts, indicates another sharp change in climate conditions at the late Pliocene–Quaternary boundary. As is known, the late Pliocene–Quaternary boundary found on continents marks a paleoclimate transition into colder and dryer climates. Data on Quaternary geology and paleontology support this assumption. The regarded region at the modern stage is in general a part of the Syrian Desert, which well manifests features of arid climate. The annual precipitation here does not exceed 200 - 300 mm. Naturally, processes of chemical weathering can not develop under such conditions.

The WC may serve as a source for some mineral deposits. For example, the mineral composition (high contents of ferric oxides and hydroxides), as well as physical-mechanical properties (dispersity, homogeneity etc.) suggests that the Syrian laterites would be a good source of raw material for mineral pigment manufacturing. In comparison to synthetic pigments, the products would be cheaper and purer ecologically.

Global changes in the environment and climate of southwestern Asia and eastern Africa during the last 200 Ma took place in the middle-late Miocene and the Quaternary. Cerling and Quadre (1990), studying sulfur and oxygen isotopes, revealed a reliable correlation between the two, and, in addition, an interrelation between the $\delta^{14}C$ value in paleosoil carbonates and the character of a biomass developed at the surface of an ancient time. The published data shows that the C_3 ($\delta^{13}C = -27‰$) biomass dominated the province 20 to 8 Ma. The expansion of the C_4 ($\delta^{13}C = -13‰$) biomass (grassy vegetation similar to that of the modern Serengeti Valleys) started between 8 and 6 Ma. The absolute substitution of the C_3 biomass by the C_4 biomass took place in all ecosystems during the last million years. An analysis of oxygen isotope behaviour in paleosoil carbonates shows a significant change in isotopic composition during the 9 - 6 Ma interval from $\delta^{18}O_{PDB}$ (−11 to −8) to (−7 to −2). This indicates a change from a dry to a humid climate, as the sediments of a dry climate are poor in heavy oxygen isotopes (Faure, 1987; Hevs, 1983). Thus, the global reorganization of atmospheric circulation is, supposedly, the main reason for changes in sulfur and oxygen isotopic ratios in paleosoils of the studied region during the last million years.

The synchronous changes in WC types, the oxygen and sulfur isotopic composition in soil carbonates, and the character of the biomass deserve attention. All this indicates that climate characteristics were very constant, and persisted during significant time intervals: a moderate and warm, relatively dry climate prevailed from 21 Ma until 7 - 8 Ma; a humid subtropic climate, from 6 Ma until 3 - 2 Ma; and a dry arid climate, from 2 Ma to the present.

It is interesting that such a conservatism of climate is generally typical of more limited time intervals, even those connected with serious climate shocks such as glaciation periods. For example, Lautenschlager (1991) demonstrated that the glaciation, which occurred 18 ka and changed energy component of the climate by 15%, did not result in dramatic changes to the global energetic cycle, thereby preserving the basic structure of atmospheric circulation. This indicates that principal changes in climate characteristics are hardly dependant on the local, even large-scale factors, like glaciations; they are rather the result of global reorganizations in the system of atmospheric circulation. According to our data, and results of the study of stable isotopes in soil carbonates, such changes were rapid, during 1 - 2 Ma, which is followed by a continuous period of stable atmospheric circulation.

The results of global changes, taking place on the land during the late Cenozoic, are reflected in the chemical composition of the seawater. Three periods of rapid increase of $^{87}Sr/^{86}Sr$ isotopic ratios (24 - 16 Ma, 5.5 - 4.5 Ma, 2.5 - 0 Ma) are recognized, and explained by sharp increases in river run-off. The latter is connected with the rapid uplift of land in the middle Miocene, the Himalayas in particular, caused by the subduction of the Indian Plate beneath Eurasia (Lauteschlager, 1991). The Greater Caucasus Range uplift in the middle-late Miocene (Borsuk *et al.*, 1989), the origin of the Anatolia–Caucasus–Elburs Andesite–Latite Volcanic Arc (Sharkov *et al.*, 1989), and the rapid subsidence in deep water depressions of the Black and Caspian Seas also deserve attention. These events are apparently interdependent; large astenosphere diapirs rising under the Red Sea Rift area caused abundant basalt volcanism in Arabia, and initiated the subduction of the Arabian Plate beneath Eurasian Plate (Sharkov and Svalova, 1991). Similar processes took place in western Europe, where the mountain systems of the Alps, the Carpathian, and the Dinarides appeared as a consequence of the Africa–Eurasia collision.

The uplift of the Tibetian Plateau, the continuing uplift of the Himalayas and the Hindu Kush, as well as the uprise of southwest Asia mountain systems such as the Zagros and the Taur, continuing to the present time (Molnar and Tapponier, 1975), were also important events at the Pliocene–Quaternary boundary. The time of these events coincides with the dramatic climate changes in the studied region. Moreover, it is very important that none of the related events took place in the Atlantic sector; where the opening of the Atlantic ocean continued.

So, the huge mountain range, extending across nearly the whole supercontinent, sublatitudinally appeared in the middle-late Miocene as a result of events in the Red Sea Rift and the collision zone between the Indian Plate and the southern part of Eurasia. Apparently, this was one of the main reasons for changes in the atmospheric circulation of the northern hemisphere, causing climatic changes throughout the vast territory. This event also influenced the hypergene processes, including the pro-

cesses of laterite formation. It is not inconceivable that the consequent principal climate change is somehow connected to the uplift of the Tibet and the Hindu Kush at the Pliocene–Quaternary boundary. The available data do not lead us to the conclusion that there was a significant long-term influence of volcanic processes on the large-scale changes in the atmosphere, however, they are an important indicator of deep-seated processes forming the surface relief of the Earth and, thereby, determining important peculiarities of atmospheric circulation.

CONCLUSION

(1) A method of numerical WC dating is proposed; and used to get results. The WC ages are estimated using isotopic data from the host rocks (basalts). The dated WC are considered indicators and reference time points of corresponding physical-geographical and paleoclimatic conditions. This concept was applied to the study of sections of diverse flood basalts in Syria [early Miocene (20.9 - 20.4 Ma), late Miocene–early Pliocene (9.7 - 4.8 Ma), and late Quaternary (0.3 Ma)].

(2) The WC of montmorillonite type were formed in Syria 21 - 9 Ma, the laterite WC of gibbsite–halloysite–goethite type, at least from 5.5 Ma to 4.5 Ma. The formation of WC ceased at the Pliocene–Quaternary boundary. The evolution of the weathering processes indicates significant changes in the integral characteristics of climate conditions in the region, from semiarid (Miocene) to humid subtropic (early Pliocene), and, finally, to the dry, arid climate of the Quaternary.

(3) Isotopic-geochronological data indicates that WC formation occurs in nearly a moment, geologically speaking (from thousands to a million years).

(4) The results obtained on the geochronology and the formation conditions of the Syrian WC reflect the evolution of geological and climate processes in the studied region, and correspond to stages of the late Cenozoic history of the Eurasian supercontinent and adjacent regions of east Africa. Significant stability of climate regimes and comparatively rapid transitions from one regime into another deserve special attention.

(5) A definite coincidence of climate change periods with periods of volcanic activity is observed; however, there is no evidence that volcanism has a direct influence on the climate. Volcanism is, rather, an indicator of deep processes in the upper mantle of the Earth, which deform the relief of the Earth's surface through geodynamic activity and, thereby, determine important peculiarities of the atmospheric circulation.

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Methodic Principles of Study of Geological Monuments of Nature in Russia

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Abstract – Monuments of Nature (MN) are regarded as well-known exemplars of natural prodigies. Geological MN represent only one class of MN. Excellent outcrop conditions, or an outstanding morphology are their necessary attributes. A subdivision of geological MN into monotypical and polytypical, their classification into types and subtypes, their ranking according to levels of importance (global, super-regional, regional, and local), and their official status (world-wide, national, and local significance) are regarded. The suggested protection categories of MN are as follows: I – a regime of especially strict protection; II – a regime of limited protection without recommendation for mass tourism; III – a regime of limited protection with recommendation for mass tourism. The descriptive order for geological MN will be illustrated by an example.

INTRODUCTION

The conception of the preservation and use of the cultural and natural heritage of Russia is presently under development. The Council for the Study and Protection of Cultural and Natural Heritage was nominated by the Presidium of the Russian Academy of Sciences following a resolution in May 1992. According to the initial conception, "the natural heritage includes natural complexes in their separate and unique forms and elements preserved in a natural, little changed state or those altered by humans" (Vedenin and Shul'gin, 1992). The geological heritage is a constituent part of the natural heritage.

The First International Symposium on the Protection of the Geological Heritage, entitled "The Earth's Memory Is in Our Hands" (Premier Symposium..., 1991), was held in 1991 in Digne, France. About 100 geological monuments, reservations, and national parks of the world were included in the preliminarily list of the global geological heritage, according to the criteria used by the Symposium. Unfortunately, the geological heritage of Russia is not yet represented on this list (Todorov, 1991). So far, the geological monuments of Russia are also absent from the unique monograph of M. Shvartsbakh (1973). At the same time, considerable work on the registration, study, and description of geological monuments is being carried out by different organizations in Russia. However, this activity has so far been inconsistent, uncoordinated, and carried out without the required preliminary work on methodic principles.

The compilation of a data base and maps of the distribution of geological MN in the territory of Russia, which has both national and world-wide significance, is currently under way at the All-Russian Institute of Geology, St. Petersburg, Russia. This work has been developing separately, and is expected to be worked out over the long term. This article is intended to present methodic principles for the study of geological MN, as worked out during the first stage of these studies, the study and description of the geological monuments of European Russia.

The term "monument of nature" was suggested in 1818 by the famous German naturalist Alexander Gumboldt. Presently, this term is being used not only in the scientific, but also in the juridical sense. For example, according to existing legislation in the Russian Federation, "certain unique natural objects and complexes, which have relic, scientific, historical, or ecological-educational significance, and which need the special protection of the state, are declared to be monuments of nature" (Laws of the Russian Federation..., 1992).

The distinctly expressed diagnostic features of one or another natural phenomenon, or results of this phenomenon, are the main attributes of MN from the viewpoint of the natural sciences. Monuments of Nature are the well-studied exemplars of natural prodigies, so they need the special protection of the state and may be used for further scientific study as well as for educational purposes. MN are the property of the state, according to the "Laws of the Russian Federation..." (1992). The objective of natural scientists is the recognition and promotion of the study of MN; the task of various levels of administrative organs is their reservation.

Geological monuments (or "geotopes" after Wiedenbein, 1992) represent one MN class; other classes include botanical, zoological, and landscape monuments, among others (Arkhipova, 1980; Atoyan, 1990). In addition to the above noted requirements, common to all MN, excellent exposure or a prominent morphology are the essential features of geological monuments. MN are defined as complex monuments (Pysin, 1983) if their features fit into different categories, e.g., geological and botanical.

B.V. Poyarkov (1988) made a general suggestion concerning "groups of natural monuments." As applied to geological objects, we are speaking of spatially related groups of geological MN, which have a common genesis and scientific, historical, and cultural significance.

We suggest referring to some MN as lost. Thus, we make a distinction between existing MN and those which have been destroyed by natural processes (land-sliding, erosion, etc.) or by anthropogenic activity, as is happening more and more often. For example, the absence of official MN status in Russia for key sections of stratigraphic subdivisions has already resulted in the elimination of numerous local and regional type sections of national importance, as well as of those related to the general stratigraphic scale, which have world-wide significance. In particular, some quarries in the Moscow region, where different units of the Moscovian, Kasimovian, and Gzhelian Stages were originally studied (in the 1920s and 1930s), have been recultivated recently, and key sections of these stages have been lost. Evidently, the loss of a type section may result in its substitution. This may result in the disappearance of Russian designations for stages from the International Stratigraphic Scale despite their justification by the long-term activity of our predecessors. Geological MN, such as the unique locality of the late Permian insects "Tikhie Gory," and the Yubileinaya Lode on the Kola Peninsula, which is unusually rich in rare minerals, were lost, in the first case, after the flooding of Kama Reservoir, and, in the second, as a result of barbarous exploitation, respectively.

An analysis of the data bank of Russia's geological MN in the European part of Russia showed that numerous geological objects, valuable for science, were either not included among the number of nominated MN, or else their importance (official status) was unwarrantedly reduced. One of the main reasons here is the absence of geoscientific criteria for determining the levels of MN significance. In this paper, we will attempt to outline in more detail the scientific basis and criteria for the recognition of geological MN and for their classification and ranking.

Two main directions defined the methodic principles of our work: (1) the specification of object classifications into types of geological MN with the recognition of their subtypes and (2) the formulation of geological criteria for ranking geological MN according to their levels of significance in order to attain some degree of objectivity in the official status definitions. We also suggested a differentiated approach to the methods of monument protection. In particular, we proved the necessity of recognizing three categories of monuments under protection. A standard scheme of description, and a form of data submission are put forth in order to establish a data base and information search system for geological MN in Russia.

CLASSIFICATION OF GEOLOGICAL MONUMENTS OF NATURE INTO TYPES AND SUBTYPES ACCORDING TO THE OBJECT PRINCIPLE

Geological monuments are divided into monotypes, which are of interest to specific geological disciplines, such as stratigraphy, paleontology, tectonics, etc., and polytypes, which are pertinent to several geological disciplines, e.g., stratigraphy and paleontology, petrography and geomorphology, etc. Application of the term "polytypical geological monument" requires the decoding of certain inner types. For example, a polytypical geological monument can be a combination of stratigraphic, paleontologic, and geomorphologic types.

Using previously published classifications of geological MN (Baratov and Novikov, 1988; Varsanof'eva and Gekker, 1951; Korotenko *et al.*, 1985; Potemkin, 1981, etc.), we recognized the following geological MN types (some of them for the first time):

1. Stratigraphic
2. Paleontologic
3. Mineralogical
4. Ore-Petrographical
5. Geochemical
6. Seismic
7. Tectonic
8. Cosmogenic
9. Geomorphologic
10. Cryogenic
11. Hydrologic-Hydrogeologic
12. Geothermal
13. Historical geological mining

To detail further the object classification of geological MN, we need to recognize not only types, but also subtypes, with the exception of geochemical and cosmogenic types, where subtype recognition is impossible due to a deficit of data. Brief characteristics of the proposed types of geological monuments and their divisions into subtypes are described below.

(1) Natural monuments of the stratigraphic type include natural or artificial outcrops of stratified formations, which bear information on age, volume, and the relations of these formations to their stratotypes, or else represent a typical sequence of beds, which are characterized by sedimentologic, biostratigraphic, paleomagnetic, and other data. We recommend referring the key and reference sections of stratigraphic subdivisions of different levels, pertinent to the general, regional, and local stratigraphic scales, to this type monument. The reference sections define stratigraphic volume, essence, and the denominations of all stratigraphic subdivisions in accordance with the principles of "The Stratigraphic Code" (Zhamoida *et al.*, 1992) and existing international practice (International..., 1976). Therefore, namely the reference sections must be considered as geological MN of the stratigraphic type.

The subdivision of geological monuments of the stratigraphic type into subtypes is quite conditional. For the Precambrian (excluding the Vendian), subtypes correspond in range to the era; for the Vendian and Phanerozoic (excluding the Anthropogene) they correspond to the series volume, and for the Anthropogene they correspond to the stratum.

(2) Natural monuments of the paleontologic type represent localities of paleontologic remains that are uniquely preserved, have a peculiar taxonomic composition, and allow the resolution of both general and specific questions concerning biology, paleontology, paleogeography, paleoclimatology, stratigraphy, and other geological disciplines. For example, the significance of vertebrate and most ancient microfossil localities are well known for their help in developing the general theory of evolution and for the study of the origin of life respectively. We first identified the following subtypes: (a) microbiotas, algae, and fungi; (b) higher plants; (c) marine invertebrates; (d) land and freshwater invertebrates; (e) vertebrates.

Geological ages are used to designate paleontological subtypes with similar scale units, just as was done for the stratigraphic subtypes, i.e., "vertebrates of the Eocene."

(3) Natural monuments of the mineralogical type are localities containing rare minerals and well-formed mineral crystals, their crystallographic and aggregate forms, unusual mineral paragenesis, and areas of obvious, current mineral-forming processes. The pertinent subtypes are: (a) localities of rare minerals and gemstones, (b) localities of well-formed crystals, druses, and mineral pseudomorphs, (c) localities of rare mineral parageneses, (d) localities of minerals with an unusual morphology, and (e) localities of present-day mineral formation.

(4) Natural monuments of the ore-petrographical type include: outcrops and massifs of standard and rare rocks, or ores with obvious details of their composition, structure, and texture; remnants of ancient volcanic apparatus; and peculiar forms of the intrusion and layering of magmatic products. The subtypes in this case are: (a) sedimentary rocks and ores, (b) magmatic rocks and ores, (c) metamorphic rocks and ores, (d) hydrothermal-metasomatic rocks and ores, and (e) weathering crusts and hypergenic ores. Either the terms "rock" or "ore" should be emphasized in the designation of ore-petrographical subtypes depending on the specifics of the concrete object.

(5) Natural monuments of the geochemical type include regions or rock massifs with anomalous contents of trace and rare elements that have scientific significance. Outcrops with clear manifestations of an iridium anomaly at the Mesozoic–Cenozoic boundary, widely discussed in recent years, may be referred to this type.

(6) Natural monuments of the seismic type are the traces left on the earth's surface and in geological section by earthquakes. The results of seismic processes are expressed in relief as gap fractures, swells, and troughs that are often flooded by underground or surface waters. Also included in this type are turbidites, olistoliths, and olistostromes in the sections. Two subtypes can be defined: (a) endogenic and (b) technogenic.

(7) Natural monuments of the tectonic type include zones and regions where well-exposed sections reveal the results of tectonic and geodynamic processes, as well as areas showing clear elements of different ranks of tectonic structures either on the continents or in transition zones between continents and oceans. For example, these are accretionary, collisional, folded, and disrupted structures, as well as others peculiar to tectonically reworked lithologic and magmatic objects. The following subtypes are suggested: (a) regions of platform structures, (b) regions of geosyncline-fold systems, (c) regions of island arcs, and (d) tectonic deformations.

(8) Natural monuments of the cosmogenic type include astroblemes and their groups and also regions of meteorite impacts containing traces of their influence on the host rocks.

(9) Natural monuments of the geomorphologic type include separate relief forms or their complexes, which most vividly reflect the interaction between endogenic and exogenic geological processes in which the exogenic dominate. The relief forms, which have specific aesthetic, cognitive, and recreational value, are included in this type. The following subtypes are recognized: (a) outliers of weathering, (b) objects of accumulative eolian relief, (c) objects of denuded eolian relief, (d) objects of erosional relief (with the exception of river terraces considered as independent subtypes), (e) river terraces, (f) objects of glacial and water-glacial relief, (g) glaciodislocations, (h) surface karst, (i) underground karst, (j) regions of abrasive coast, (k) regions of accumulative coast, (l) objects of land-slide relief, (m) volcanoes, (n) mud volcanoes, (o) mountains-laccoliths, (p) objects of cryogenic relief, and (q) tectonic slopes. This list is not all-inclusive and may be enlarged in the future.

(10) Natural monuments of the cryogenic type are specific cryogenic formations that are typical of permafrost, with the exception of those formations expressed in relief and related to the previous type. The subtypes are: (a) ice sheets, (b) ice-derived formations, (c) bodies of underground ice, (d) cave ice, and (e) relics of permafrost.

(11) Natural monuments of the hydrologic-hydrogeologic type include rare and unique natural groundwater discharges (springs) and reservoirs of surface water (lakes). The intimate connection between underground and surface waters explains their placement in one geological MN type. In this case, the main criteria for the selection of ground-water sources as geological MN are the output volume and the gas-hydrochemical and physical properties of the waters. Lakes that are considered to be natural monuments, like the springs, must be unique in the gas or chemical composition of their waters, or in the properties of their natural bottom deposits, hydrological regime, etc. MN of this type are divided into the following subtypes: (a) springs with an unusual output volume, (b) sources of mineral water, (c) geysers, (d) lakes with a specific hydrological regime, and (e) lakes with a specific hydrochemical regime.

(12) Natural monuments of the geothermal type include areas of unique and rare natural geothermal events on the Earth's surface. Outflows of dry hot gases, water streams, and thermal water onto the day surface are often characteristic of these areas. Two known subtypes are: (a) exogenic and (b) endogenic. Occurrences caused by hypergenic processes in the Earth's interior are referred to the exogenic subtype. Yangantau Mountain in Bashkortostan is an example of this subtype. The endogenic subtype may include geothermal occurrences in regions of young and recent volcanism, e.g., Goryachii Beach on Kunashir Island, the Pauzhatka Springs in Kamchatka, etc.

(13) Natural monuments of the historical geological mining type are the monuments of mining history of a certain territory. The subtypes are: (a) monuments of the history of geology and (b) monuments of the history of mining exploration. The latter include not only monuments of nature, but of technology too.

ARRANGEMENT OF GEOLOGICAL MONUMENTS OF NATURE ACCORDING TO THEIR LEVEL OF SIGNIFICANCE AND OFFICIAL STATUS

In our ranking of the geological MN by level of significance, we paid particular attention to geological criteria. There are four such levels of significance: global, superregional, regional, and local (the table). Definitions of these significance levels for different types of geological monuments, based on selected criteria, are presented in the table. The ranking of some types of geological monuments was not done due to the lack of study of available materials on these monuments. It was impossible to distinguish between the levels of significance for geological monuments of the historical geological mining type, because their rank is determined by historical, rather than geological criteria. All four of the given significance levels can be defined only for geological MN of the stratigraphic, paleontologic, petrographic, and cosmogenic types. It is difficult to recognize global level monuments among MN of the tectonic type, because the scale of global tectonic events does not fit the concept of a monument, which must have limited dimensions. We used the system of Krasnyi's tectonic terms (1991, 1992) to rank natural monuments of the tectonic type.

In addition, not every significance level was recognized for MN of the geomorphologic and hydrologic-hydrogeologic types. There are no natural monuments of the geomorphologic type on the global level, because the relief forms reflect general regularities in the development of the Earth's planetary shells, continents, and their main parts cover vast territories, and, once again, do not satisfy the MN concept. Therefore, among those monuments of the geomorphological type, only groups of monuments, rather than isolated ones, can be characterized on a super-regional level. As for monuments of the hydrologic-hydrogeologic type, it is difficult to distinguish them on super-regional and regional levels. This may be explained both by the global distribution of natural waters and by difficulties in ranking the corresponding factors of the formation of this MN type, such as climate, geology, geomorphology, and other phenomena. Aside from the significance level criteria of geological MN considered in the table, it is necessary to consider a number of additional conditions that allow a concrete geological monument to rise to a level of significance. Such conditions include: (a) the combination of several lower ranking monument types into a single MN, (b) the occurrence of any rare feature within a given territory, and (c) the epitomizing expression of any given type-feature of a monument under consideration as compared to other expressions of this feature.

The official status of a geological monument, i.e., its classification as a MN of local, national (or republican, or federal, etc., since a generally accepted nomenclature is currently absent), or world significance, is to be defined according to its significance level (see above). These relations will be approached as follows: MN on the local level have localized significance; those of national level have regional and super-regional importance; monuments on the global level have global importance. Meanwhile, monuments showing rare geological phenomena, e.g., those of the cosmogenic type, always have great scientific and cognitive value. Therefore, the status of such natural monuments should be higher. For example, MN of the cosmogenic type on the local significance level (according to geological criteria) may be given a MN status on the national level.

The official status of MN, when they have educational, aesthetic, and recreational value in addition to scientific importance, may increase as a result of: (1) the object's well-expressed and typical geological features, which allow even the inexperienced geologist to understand its most essential peculiarities, (2) the beauty of the landscapes of which the geological monument is a part, and (3) easy access to the monument, because of available highways or railroads, which allow it to be included on a list of regional, scientific, touristic, or other points of interest.

CATEGORIES OF GEOLOGICAL MONUMENTS OF NATURE TO BE PROTECTED

Determining which categories of natural monuments will be protected depends mainly upon the current economic situation in Russia, especially with its entrance into a market economy. A significant number of geological MN, in particular the paleontological and mineralogical types, became objects of extensive, commercial activity. Absence of regulations, with regard to such activity with the geological MN, has already brought about the continuing irreplaceable loss of such monuments, many of which are unique on the global scale. That is why we believe it is necessary, as a preventive measure, to distinguish between three categories of geological MN protection.

Arrangement of various types of geological monuments of nature according to their levels of significance (using geological criteria)

Level of significance according to geological criteria	General characteristics	Type of monument		
		stratigraphic	paleontologic	petrographic
Global	Objects that characterize general regularities in the development of the planetary shells and main heterogeneity of the Earth, and which are unique on a global scale	Typical sections and areas where general stratigraphic subdivisions, accepted on a global scale, have been justified	Locations of paleontologic remains, which allow us to solve general biologic questions concerning the origin and evolution of life and the development of large taxonomic subdivisions (superkingdom, kingdom, type, division, class) to be answered	Localities of effusive and metamorphic rocks with type significance on a global scale, their occurrences that reveal the magmatic and metamorphic processes of the lithosphere class; outcrops of rare types of sedimentary rocks (for example, those unique for their use in determining the stages of geological history)
Superregional	Objects reflecting superregional regularities in the development of the continents, oceans, and transition zones	Stratotypes of general stratigraphic subdivisions accepted in the former Soviet Union	Localities of paleontologic remains unique in their taxonomic compositions or in their state of preservation, which allows us to obtain information about the structure of these paleontologic objects or their parts (biomineralization, etc.), the taxonomic diversity of large biochores, the general structure of the paleobiocoena, and their biogeographic differentiation	Outcrops and massifs that characterize definite types of geological formations
Regional	Objects reflecting the processes of development of large regions (processes that are either typical or rare and unique for given regions)	Stratotypes of regional stratigraphic subdivisions	Localities of well preserved paleontologic remains or of taxonomic assemblages unique in the region or reflecting paleobiocoena of various biofacies, which provided the basis for regional stratigraphic subdivisions	Outcrops and rock massifs in which the processes of regional rock formation were most vividly manifested
Local	Objects reflecting processes that are typical of parts of regions	Stratotypes of local stratigraphic subdivisions	Localities with well-preserved paleontologic remains or with various taxonomic compositions of different biofacies, as well as the reference sequences of local stratigraphic subdivisions	Outcrops and rock massifs that are either typical or rare in the region

Type of monument			
tectonic	cosmogenic	geomorphologic	hydrologic-hydrogeologic
?	Manifestations of large-scale impact events of the geological past, which may have influenced the outer shells of the Earth on a global scale, i.e. the giant astroblemes with diameters of up to 50 - 100 km and more; beds bearing traces of such events in the form of mineralogical, geochemical, lithologic, and paleontologic features	Not distinguished	Springs with a unique yield; springs and lakes containing water unique in its mineralization, chemical composition, physical properties, or having bottom deposits unique in their composition or properties
Objects reflecting stages of the tectonic evolution of large parts of continents, oceans, or transition zones: platforms, fold belts, marginal seas, and the most important crustal and crust-mantle faults	Astroblemes of medium size (10 - 50 km), indicating events that may have had superregional consequences for the outer shells of the Earth	Large relief forms reflecting the processes of development of continental platforms and fold belts, or of oceanic and transitional structural zones. In size, these correspond not to separate monuments of nature, but to their groups	Springs with a rare gross yield; springs and lakes with water or bottom deposits having a rare chemical composition or degree of mineralization, or unique physical properties; springs with unusual structure and regime, or with water outlets (mud volcanoes, geysers) with a unique character
Objects reflecting the most typically manifested structures that are typical for regions: fold systems, rifts, ophiolite sutures, accretion-collision belts, geoblocks, etc.	Lesser astroblemes (3 - 10 km), the influence of whose corresponding impact events may have had regional significance	Complexes of relief forms reflecting regional processes of relief formation, or else rare forms of reliefs and their combinations	
Various structures manifested in relief or in outcrops: systems of folds, domes, olistostromes, nappes, typical faults (strike-slip, normal, etc.)	Small astroblemes (less than 3 km) and meteorite craters and their groups	Small or separate relief forms having scientific, educational, or aesthetic significance	Springs with a high yield for their area; springs and lakes whose water or bottom deposits have (for a given area) unique chemical compositions or physical properties; springs and lakes valuable in scientific, educational, and aesthetic respects along with those significant as standards, for example, in the quality of their water

The first category requiring especially strict protection comprises those objects that are valuable for either commerce or sample collection. Here we are speaking mainly of localities with rare examples of paleontological remains, e.g., vertebrates, cephalopods, insects, etc., as well as gemstones and mummies. MN in this category are mostly subject to the risk of being completely plundered and destroyed. These require a special regime of protection, even forbidding of the disclosure of their locations in printed publications, as well as a limitation on sample collecting (sampling may be carried out only by specialists after getting permission from the organization in charge of their protection).

The second category of protection (limited protection without recommendation for mass tourism) includes geological monuments with special scientific importance. Such monuments may be the objects of geological excursions, field trips for the education of students, and sample collecting without special permission. These MN are not interesting from an aesthetic and recreational point of view, and are not desirable for mass tourism. Characteristic examples from this category include the stratigraphic and cosmogenic types of geological MN, which are not extremely attractive, though they are important only to specialists and the occasional amateur. Information on the locations of these MN may be published in the open press.

The third category of protection (limited protection with a recommendation for mass tourism) covers the monuments important for tourism and study. MN of this category are either picturesque (attractive rocks, grotto, caves, lakes, springs, etc.), or else are valuable for public education, e.g., having visible manifestations of geological processes, localities with mass accumulations of widely distributed paleontological remains, monuments to the history of geology and mining, etc. Numerous geomorphological MN (karst caves, mud volcanoes, mountain-laccoliths, and erosive relief forms) and outwardly attractive MN of the ore-petrographic (rocks) and tectonic types belong to this category. MN in this group should be readily accessible, but the remoteness of some such monuments (for example, their location on a polar island) may make them appear more exotic. Monuments in this category may be used for the following purposes of tourism: attracting tourists (including foreigners) and selling maps, booklets, post cards, badges, souvenirs, etc.

WAYS TO DESCRIBE GEOLOGICAL MONUMENTS OF NATURE

The description of geological MN may be carried out according to the following suggestions:

The monument or group of monuments should be designated along with the indication of its synonyms; if the monument has been lost, this should be indicated in parentheses after the designation.

The location of the monument should be given with the administrative region and the topographic coordinates: nomenclature of the map sheet with scale 1 : 200000 and the position mark on maps with scale from 1 : 1000000 to 1 : 5000000.

(I) Brief characteristics of the monument (compared with known world and national analog, if possible): the history of investigatory, scientific, and training excursions to the monument; the places in which the most representative and standard collections are stored (in this case, the location of gemstones, preserved artifacts, and constructions made from them should be pointed out) – 0.5 - 1.0 pages.

(II) The condition of the monument and necessary measures for its preservation.

(III) The type and subtype of the monument: for polytype monuments, the combination of all manifested types and subtypes should be indicated.

(IV) The level of significance of the monument according to geological criteria.

(V) The official status of the monument and the form of its incorporation into the natural-reserve foundation of the Russian Federation: the monument's location in the territory of a state natural park, a national natural park, a resort area, or a medical zone; the present and proposed status of the monument (world, national, or local importance); the number and date of the resolution on its protection; and the state organization, which approved the protected status of the monument.

(VI) Protection category.

Note. Signs or symbols that allow the MN to be counted as a complex phenomenon should be present (apart from the geology, it may be the botany, landscape, etc., which makes the MN significant); the religious, historical, and cultural significance of the monument should also be noted.

Publications (bibliography of all cited works).

A photograph of the monument (if possible).

The author of the description and his or her signature.

AN EXAMPLE OF A DESCRIPTION OF A GEOLOGICAL MONUMENT OF NATURE

Cambrian and Ordovician Deposits of the Sablinka River (syn: the waterfall on the Sablinka River and caves on the banks of the Tosno River within the limits of the Ul'yanovka settlement; Sablinka Ecological-Geological Reservation).

St. Petersburg region, Tosno District; on the banks of the Tosno River (below the village of Gertovo) and its right tributary, the Sablinka River; 0-46-1.

(I) Outcrops of this monument have been known since the beginning of the nineteenth century; ideas concerning the geology and stratigraphy of the Cambrian and Ordovician of the Russian Platform were first developed here, and some well-known paleontological remains were first described in this area. Stratotypes of the Sablinka Formation of the middle Cambrian and the Tosno Formation of the lower Ordovician are represented in this monument's sections.

Quartz sandstones and sands of the Sablinka Formation lie upon lower Cambrian blue clays. Sands of the upper Cambrian Ladoga Formation, with remains of inarticulate brachiopod shells, conodonts, and problematics (*Rukhinella*), occur higher up in the sequence. The Tosno Formation is represented by sands containing similar organic remains, but with a different taxonomic composition. Argillites, sands, and aleurites of the Kopora Formation and glauconite sands of the Leeta Formation (Latorpian Horizon), attributed to the lower Ordovician, terminate this section.

This monument is also interesting from the mineralogical point of view, because of the marcasite pseudomorphs occuring here, which comes after pyrite and vice versa (pyrite is a common mineral in Cambrian clays, though marcasite is not); there is also coarse-crystalline calcite (anthraconite) occuring here in the form of monocrystals and druses.

The monument was the object of excursions by the First All-Russian Geological Congress in 1922 and the 27th International Geological Congress in 1984. It serves also as an area of annual training excursions for students of St. Petersburg University, the St. Petersburg Mining Institute, and the A.I. Gertsen Russian State Pedagogical University. The representative fauna samples collected along the Sablinka River are kept at the F.N. Chernyshev Central Scientific-Research and Geological-Prospecting Museum in St. Petersburg and at the Institute of Geology of the Estonian Academy of Sciences in Tallinn.

(II) The condition of monument is satisfactory.

(III) This is a polytype monument: there is a combination of stratigraphic (lower Cambrian–lower Ordovician), mineralogical (a locality of well developed crystals, druses, and minerals pseudomorphs), and historical-geological-mining (geological monument of history) types.

(IV) Level of significance – superregional.

(V) A monument of national importance: it was authorized as a monument of local importance by resolution no. 145 of the Executive Committee of the Leningrad Regional Council on March 29, 1976; then received its national status as Sablinka Ecological-Geological Reservation by resolution no. 98 of the Lesser Council of the Leningrad Regional Council of People Deputies on June 17, 1992.

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CONCLUSION

Only the initial results of scientific and methodic studies on the geological monuments of nature in Russia were reported in this paper. We appeal to all geologists, specialists in other fields, regional specialists, and amateur naturalists to send to the All-Russian Institute of Geology in St. Petersburg all their data concerning geological MN that have scientific importance and need protection. These descriptions should be submitted according to the above format.

ACKNOWLEDGMENTS

In conclusion, we would like to express our deep gratitude to R.L. Brodskaya, P.I. Zhamoida, L.I. Krasnyi, V.L. Masaitis, and V.V. Starchenko for their advice, helpful contributions, and discussion of these materials.

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Triassic Bivalve and Gastropod Assemblages from Western Kazakhstan and Their Stratigraphical Significance

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Abstract – The bivalve and gastropod assemblages are described and used for correlation of the lower and middle Triassic stratigraphic sequences of the Pricaspian, Mangyshlak, and Ustyurt.

Triassic deposits in the Caspian Sea depression, Ustyurt, and Mangyshlak are represented principally by brackish-water and continental sediments. Their great facial variability and the rarity of their paleontological remnants compelled us to apply biostratigraphy and correlation of any fossils: Ammonoidea, bivalves, gastropods, conchostracan, ostracods, etc. Because the Ammonoidea occur only in the Olenekian Stage of some regions, bivalves and gastropods, which have relatively high rates of evolution, take on a leading role. They are widespread in coastal-marine and shallow-water facies where they form large accumulations, and are found much more frequently than Ammonoidea. As bottom organisms, they are of great interest not only to stratigraphy but also to paleogeography. As a result, the available classical and contemporary works devoted to the study of the Triassic bivalves mainly from the northern Mid-Siberia (Dagis *et al.*, 1985) and the northeastern Russia are of great significance (Bychkov *et al.*, 1976; Vozin *et al.*, 1964). At the same time the Triassic bivalves of the southern regions are insufficiently studied. This concerns, in particular, the territory of western Kazakhstan. Until recently, no special attention was paid to gastropods due to difficulties involved in their preparation and poor preservation. However, the available paleontological data on gastropods derived from the stratotype sections of the Mangyshlak mountain range, and from the drilling cores recovered in southern Mangyshlak, Ustyurt, and the Caspian Sea region indicates (similar to the bivalves) that the majority of their forms are easily classified up to the species level. Thus they provide a reliable stratigraphic basis.

The bivalves and gastropods in the Triassic sequence of western Kazakhstan are distributed irregularly. This fauna is most abundant in the Olenekian sediments of the Mangyshlak and the Caspian Sea depression. Its depleted assemblages are found in the middle Triassic sediments of the Mangyshlak, Ustyurt, and the Caspian Sea region. Only bivalves were recognized in the upper Triassic sediments of the Mangyshlak mountain range. It is quite evident that this distribution pattern is controlled by specific conditions, which were present in the Triassic paleobasins, and are considered to have had the following environment.

At the Olenekian Stage a marine basin ingressed from the Tethyan region to the Caspian Sea depression. It occupied the territory of the Eastern Ciscaucasia, Tuarkyr, Mangyshlak, and the southern part of the Caspian Sea depression. The connection with the Tethys Ocean ceased in the pre-Anisian time, which is evidenced by the fact that there is no Ammonoidea younger than Olenekian in the sequences. The middle Triassic time was characterized by brackish-water basins permanently connected with each other. Alluvial and lacustrine-marsh conditions existed in the Caspian Sea region, Ustyurt, and southern Mangyshlak during the late Triassic, when marine conditions retained only in the Mangyshlak mountain range.

Certain similarities between the compositions of gastropod and bivalve assemblages point to their origin in a single paleobasin. This allowed us to use them for the correlation of the Triassic sequences of Mangyshlak, Ustyurt, and the Caspian Sea depression. Moreover, some of their species, which have a narrow vertical range of distribution, served as a means of dating the corresponding sediments.

The oldest of known bivalve assemblage in southern Mangyshlak is related to the Tartali (1100 m thick), Karadzhatyk (up to 600 m thick), and Fetisovo formations (600 m thick), which are made up of mudstone and siltstone with interlayers of limestone, sandstone, and dolomite. The assemblage is represented by species *Bakevellia pannonica bogdoensis* Kipar., *B. kambei* Nakaz., *Eumorphotis multiformis* (Bittn.), *Entolium microtis* (Witt.), *Mytilus tuarkyrensis* Kipar., *Modiolus triquetrus curtus* Ench., *Ornithopecten temirbabensis* Kipar., *Heminajas* (?) *balatonis* (Frech), and *Leptochondria minima* (Kipar.) (Fig. 1). The majority are typical only of the Olenekian sediment. Among the forms wider in stratigraphic range, *Bakevellia exporrecta* (Leps.), *B. costata* (Schloth.), *Gervillia mytiloides* (Schloth.), *Unionites fassaensis* (Wissm.), and *U. canalensis* (Cat.) were found in the lower and middle Triassic deposits, and *Pleuromya musculoides* Schloth., *Neoschizodus laevigatus* (Ziet.), *Palaeoneilo*

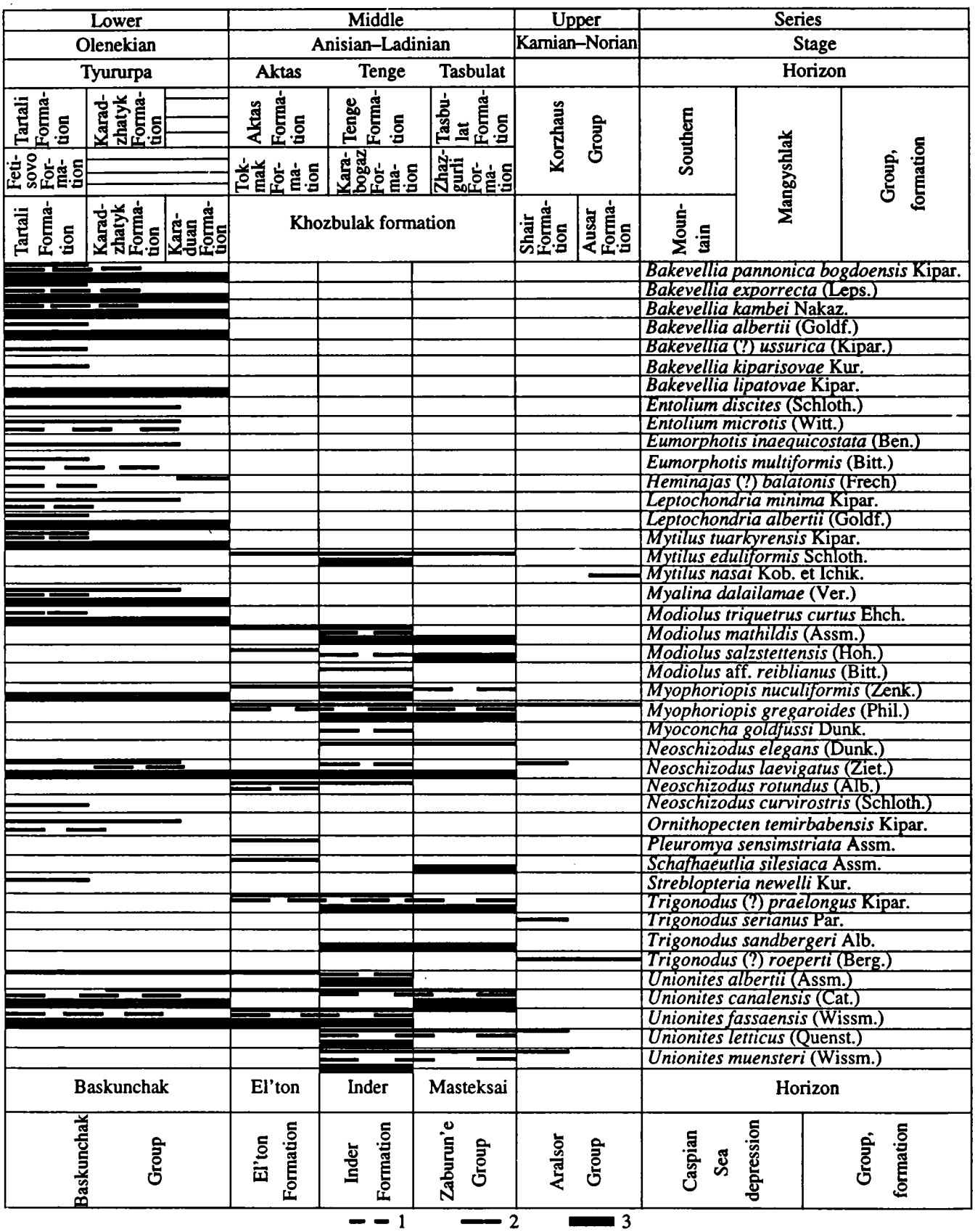

Fig. 1. Stratigraphic distribution of bivalves in the Triassic deposits of Western Kazakhstan.

Species found in the Triassic deposits of: 1 – southern Mangyshlak, 2 – the Mangyshlak mountain range, and 3 – the Caspian Sea depression.

elliptica (Goldf.), and *Leptochondria albertii* (Goldf.) were recognized in the lower to upper Triassic strata.

The lower Triassic bivalve assemblage is much more diverse in the Mangyshlak mountain range. *Edentula castelli* (Witt.), *Myalina dalailamae* (Ver.), *Bakevellia* (?) *ussurica* (Kipar.), and *Eumorphotis inaequicostata* (Ben.) were found here apart from the above mentioned forms with a narrow stratigraphic range. The number of accompanying species also increases: *Streblopteria newelli* Kur., *Entolium discites* (Schloth.), *Bakevellia albertii* (Goldf.), *Unionites albertii* (Assm.), and *Neoschizodus curvirostris* (Schloth.) were noticed in addition to the already listed species. This assemblage characterizes the Tartali, Karadzhatyk, and Karaduan Formations (450, 400, and 270 m thick respectively), which are composed of siltstone and mudstone with interlayers of limestone and sandstone.

The bivalves, definable down to the species level, have not yet been found in the Ustyurt.

The most complete lower Triassic bivalve assemblage in the Caspian Sea depression was recovered in Bol'shoe Bogdo Mountain from the Baskunchak Group sequence represented by mudstone and siltstone with limestone interlayers. It is 260 m thick (Kiparisova *et al.*, 1972). The rocks are somewhat less rich in fossils than the above described assemblages. As in the Mangyshlak, the assemblage consists of the species with a narrow stratigraphic range: *Mytilus tuarkyrensis* Kipar., *Myalina dalailamae* (Ver.), *Bakevellia pannonica bogdoensis* Kipar., and *B. lipatovae* Kipar., along with forms of a wider stratigraphic range: *Unionites fassaensis* (Wissm.), *U. canalensis* (Cat.), *U. albertii* (Assm.), *Myophoriopis nuculiformis* (Zenk.), *Neoschizodus orbicularis* (Bronn.), *N. laevigatus* (Ziet.), *Parallelodon beyrichi* (Stromb.), *Gervillia mytiloides* (Schloth.), *Bakevellia albertii* (Goldf.), and *Leptochondria albertii* (Goldf.). I also defined the Olenekian *Mytilus tuarkyrensis* Kipar., *Bakevellia pannonica bogdoensis* Kipar., *B. kambei* Nakaz., and *Modiolus triquetrus curtus* Ench., as well as the lower-middle Triassic *Unionites fassaensis* (Wissm.), *U. canalensis* (Cat.), and *Bakevellia exporrecta* (Leps.) in the cores of recently drilled boreholes in the southern part of the basin.

Thus, the bivalve assemblages found in the lower Triassic deposits of western Kazakhstan are characterized by the dominance, both in diversity and abundance, of the species of the genera *Bakevellia*, *Eumorphotis*, *Mytilus*, and *Entolium*. Some endemic forms, like *Mytilus tuarkyrensis* Kipar., *Myalina dalailamae* (Ver.), *Bakevellia pannonica bogdoensis* Kipar., and *Ornithopecten temirbabensis* Kipar., are of considerable importance. The analysis of the assemblages allowed us to date the host deposits to the Olenekian, and to correlate, stratigraphically, the Baskunchak Horizon of the Caspian Sea depression to the Tyururpa Horizon of Mangyshlak.

The lower Triassic gastropods of western Kazakhstan have so far been found only in Mangyshlak. However, a review by L.D. Kiparisova on gastropods of the USSR (Kiparisova, 1973) refers to representatives of the *Neritaria* genus from the lower Triassic of the Caspian Sea depression (the Bol'shoe Bogdo Mountain and Inder Lake). I defined *Natica minima* Assm., *N. stanensis* Pich., *Neritaria subincisa* Kittl., *N. lendzinensis* Ahl., *Omphaloptycha schmidti* Assm., *Worthenia excelsior* Assm., *W. michaeli* Ahl., and *Ampullina silesiaca* Ahl. from the Tartaly, Karadzhatyk, and Karaduan Formations of the Mangyshlak mountain range (Fig. 2).

For a long time, the Karaduan Formation was attributed to the middle Triassic based on findings of bivalves and gastropods *Neoschizodus laevigatus* (Ziet.), *N.* ex gr. *elegans* (Dunk.), *Bakevellia* ex gr. *exporrecta* (Leps.), *Brochidium* aff. *anatolicum* Arth., and *Marmolatella* sp. in the Karatauchik Ridge, as well as *Myoconcha gastrochaena* Gieb., *Neoschizodus laevigatus* (Ziet.), *N.* cf. *ovata* Goldf., *N.* aff. *germanica* (Hoh.), *N.* ex gr. *elegans* (Dunk.), *Bakevellia albertii* (Goldf.), *Unionites fassaensis* (Wissm.), *U.* cf. *canalensis* (Cat.), *Naticella* cf. *costata* Muenst., and *Worthenia* sp. in the area of the Karashek Mountain (Astakhova and Mstislavskii, 1973). The above listed assemblage is insufficient evidence that the Karaduan Formation is of middle Triassic age for two reasons: first, most of the forms were classified by open nomenclature, and, second, the assemblage consists of species typical of both the lower and middle Triassic. The only exception is the gastropod *Naticella costata* Muenst., which is characteristic of the upper lower Triassic.

Later investigations (Lipatova *et al.*, 1984) established the lower Triassic (Olenekian) age of the Karaduan Formation. The lower Triassic bivalves and gastropods *Heminajas balatonis* (Frech), *Natica minima* Assm., *Neritaria subincisa* Kittl., *Omphaloptycha schmidti* Assm., *Worthenia excelsior* Assm., and *W. michaeli* Ahl., which I defined from the formation sequence, were of considerable importance in reaching this conclusion.

The stratigraphic position of the Karaduan formation accords well with the fact that one of the most complete sequences of the marine upper Olenekian in the world is situated in the Mangyshlak mountain range (Rostovtsev and Zhamoido, 1984).

Among the middle Triassic bivalve and gastropod assemblages known in the regions under consideration, the most representative one belongs to the Khozbulak Formation of the Mangyshlak mountain range. The formation is made up of limestone and mudstone with sandstone interlayers, with a total thickness of up to 1140 m. Many researchers (Astakhova *et al.*, 1973; Kiparisova, 1977) consider this formation to be of the late Triassic age. My findings of bivalves and gastropods indicate that it is of middle Triassic age, with the exception of the upper subformation in eastern Karatau containing no fossils. Bivalve species of the middle Triassic from the Khozbulak Formation are represented by *Modiolus salzstettensis* (Hohen.), *M. mathildis*

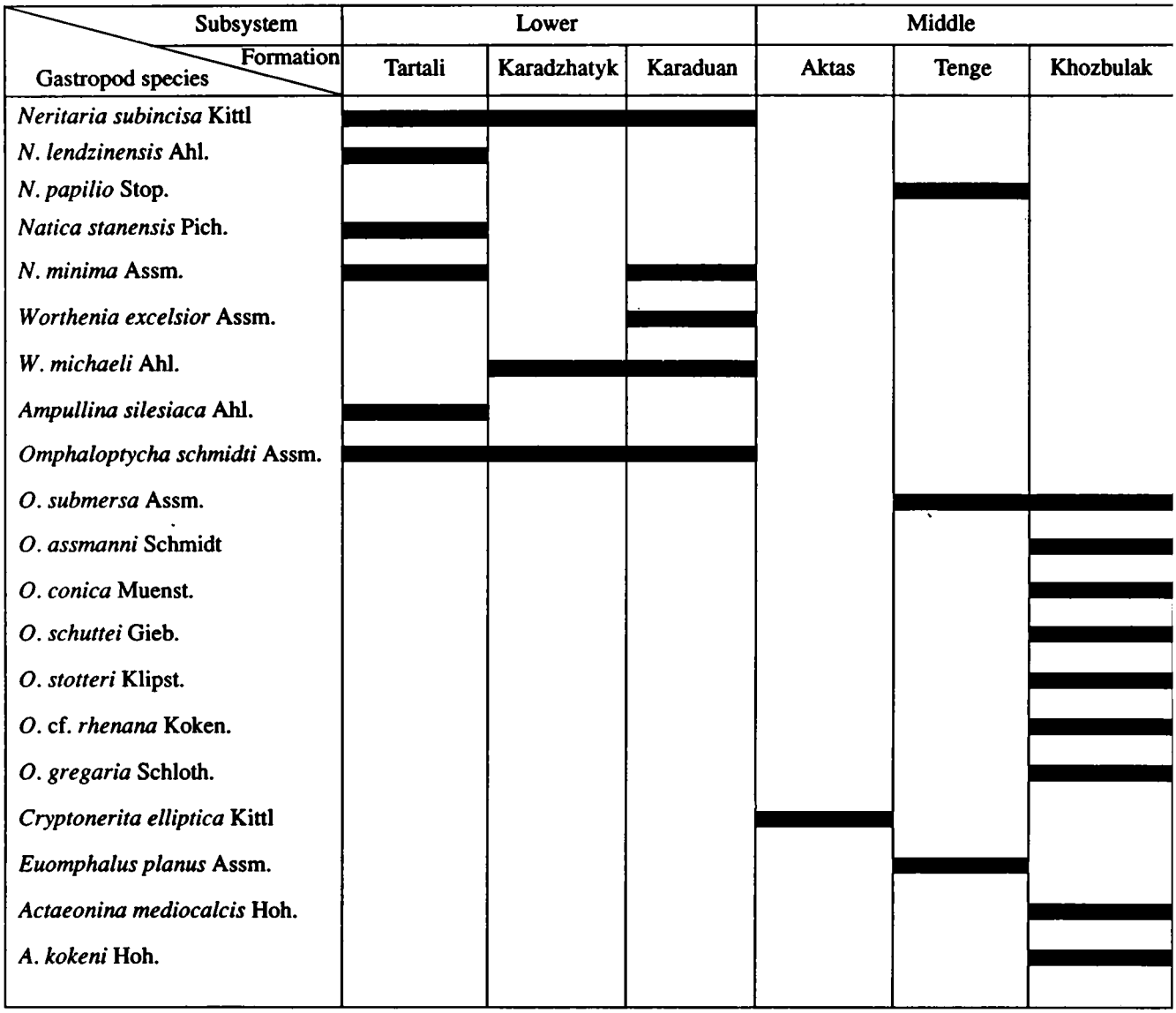

Fig. 2. Stratigraphic distribution of gastropods in the Triassic deposits of Mangyshlak.

(Assm.), *Pleuromya sensimstriata* Assm., *P. brevis* Assm., *Schafhaeutlia silesiaca* Assm., *Trigonodus* (?) *praelongus* Kipar., and *Unionites subrectus* (Bittn.). The gastropod species are represented by *Omphaloptycha conica* Muenst., *O. schuttei* Gieb., *O. submersa* Assm., *O. assmani* Schmidt., *O. stotteri* Klipst., *O.* cf. *rhenana* Koken., *O. gregaria* Schloth., *Cryptonerita elliptica* Kittl., *Actaeonina mediocalcis* Hohen, and *A. kokeni* Hohen. Together with them, bivalves, overcoming the boundary between the early and middle Triassic, were found: *Myophoriopis nuculiformis* (Zenk.), *Unionites canalensis* (Cat.), *U. albertii* (Assm.), *U. fassaensis* (Wissm.), *Palaeoneilo elliptica* (Gold.), *Mytilus eduliformis* Schloth., *Neoschizodus laevigatus* (Ziet.), and *N. elegans* (Dunk.), as well as other forms present in the middle and late Triassic: *Modiolus* aff. *reinblianus* (Bittn.), *Unionites muenstery* (Wissm.), *U. letticus* (Quenst.), *Myophoriopis gregaroides* (Phil.), and *Neoschizodus rotundus* (Alb.).

The middle Triassic bivalve assemblage of southern Mangyshlak was revealed in the Aktas, Tenge, Tasbulat, Tokmak, and Zhazgurly Formations. They are up to 140, 300, 180, 200, and 350 m thick, respectively, and are represented by limestone, dolomite, mudstone, and siltstone. In specific composition, the assemblage is analogous to that from the Mangyshlak mountain range, although it is not so abundant. Among the guide fossils in the southern Mangyshlak assemblage, there are also representatives of such species as *Trigonodus* (?) *praelongus* Kipar., *Modiolus salzstettensis* (Hohen.), and so on, while *Unionites canalensis* (Cat.), *U. fassaensis* (Wissm.), *U. letticus* (Quenst.), *U. muensteri* (Wissm.), *Myophoriopis gregaroides* (Phil.), *Myoconcha goldfussi* (Dunk.), *Gervillia mytiloides*

(Schloth.), and *Neoschizodus laevigatus* (Ziet.) are present as accompanying species.

The middle Triassic gastropod assemblage from southern Mangyshlak are significantly less rich than those from the Mangyshlak mountain range both in species diversity and abundance. Only rare representatives of *Cryptonerita elliptica* Kittl., *Euomphalus planus* Assm., and *Omphaloptycha submersa* Assm., known from the middle Triassic of western Europe, are present in this assemblage.

In Ustyurt region, rare remains of the bivalves *Unionites canalensis* (Cat.), *U. albertii* (Assm.), and *U. fassaensis* (Wissm.) were encountered in the middle Triassic part of the terrigenous sequence in the Komsomol'skaya and Horlyk exploration areas. The stratigraphic range of these species, as was already noted, is early-middle Triassic.

In the Caspian Sea depression (the Aralsor and Inder exploration areas), Kiparisova found a rich bivalve assemblage from the Inder Formation, up to 200 m thick, represented by mudstone, limestone, and marlstone. The assemblage includes the species *Neoschizodus ovatus* (Goldf.), *Myophoriopis planus* Hohen., *M.* cf. *subundatus* Schaur., and *Pleuromya pulchra* Assm., which are typical only of the middle Triassic, and *Myophoriopis nuculiformis* (Zenk.), *Parallelodon beyrichi* (Stromb.), *Gervellia mytiloides* (Schloth.), *Mytilus eduliformis* Schloth., *Unionites fassaensis* (Wissm.), and *U. albertii* (Assm.), which are typical of the lower and middle Triassic.

The paleontological data I gathered from the cores of new wells drilled in the Ural–Volga, and Ural–Emba interfluve areas allowed me to not only substantively extend the above list of bivalves, but also to reveal for the first time a gastropod assemblage. For example, in the Inder Formation and Zaburun'e Group I defined several middle Triassic species of bivalves: *Trigonodus* (?) *praelongus* Kipar., *Modiolus mathildis* (Assm.), *M. salzstettensis* (Hohen.), and *Schafaeutlia silesiaca* Assm. as well as *Trigonodus sandbergeri* Alb., *Unionites letticus* (Quenst.), *U. muensteri* (Wissm.), and *Myophoriopis gregaroides* (Phil.) existed in the middle and late Triassic. *Neoschizodus laevigatus* (Ziet.) and *Pleuromya musculoides* Schloth., identified in all three Triassic series, were also defined. The gastropod assemblage recognized so far is not very diverse and consists of the species *Omphaloptycha schuttei* Gieb., *O. gregaria* Schloth., *Actaeonina mediocalcis* Hohen., *Cryptonerita elliptica* Kittl., and *Euomphalus planus* Assm. which are widespread in the middle Triassic of western Europe.

From the aforesaid, it may be assumed that the bivalve assemblages from the middle Triassic are very similar, and include both the forms of a wide stratigraphic range (lower-middle Triassic and middle-upper Triassic) and the species limited in distribution to the middle Triassic only. In general, the assemblages discussed above are characterized by widely occurring species of the genera *Modiolus*, *Trigonodus*, *Myophoriopis*, *Unionites*, and *Pleuromya*, which allows us to date the host deposits to the middle Triassic. The assemblages of the middle Triassic gastropods are more monotonous, as the representatives of the genus *Omphaloptycha* generally predominate, while the other genera *Cryptonerita*, *Euomphalus*, *Actaeonina*, and *Neritaria* have a less significant abundance. The bivalve and gastropod assemblage peculiarities outlined above afford us the opportunity to correlate the middle Triassic deposits of Mangyshlak, Ustyurt, and the Caspian Sea depression.

In the upper Triassic deposits, only bivalves, and only in the Mangyshlak mountain range, were found in the Shair (up to 1800 m thick) and Ausar formations (400 m thick), both made up of siltstone, mudstone and sandstone with rare limestone interlayers. This is, compositionally, the poorest assemblage in comparison to those discussed above. However, its forms: *Trigonodus serianus* Par., *T. roeperti* (Berg.), and *Mytilus nasai* Kob. et Ichik., allow us to confidently correlate the host deposits with the upper Triassic of the general stratigraphic scale. In addition to the above listed forms, *Unionites letticus* (Quenst.), *U. muensteri* (Wissm.), *Myophoriopis gregaroides* (Phil.), *Neoschizodus laevigatus* (Ziet.), and *Palaeoneilo elliptica* (Goldf.) were also noted.

Thus, the discussed data on the bivalves and gastropods from the Triassic of Western Kazakhstan readily illustrates, in spite the fact that they can characterize only such units as stages or as subsystems, their importance in:

– the dating and correlation of sequences in accordance with the general stratigraphic scheme (the age of the Karaduan and Khozbulak Formation was defined more precisely, and the age of the Ausar, Tenge, Tokmak, Zhazgurly, and Fetisovo Formation was substantiated) (Zhidovinov, 1985);

– the correlation of sequences remote one from another (the Baskunchak Horizon of the lower Triassic from the Caspian Sea depression and the Tyururpa Horizon from Mangyshlak, as well as the middle Triassic deposits of Mangyshlak, Ustyurt, and the Caspian Sea depression were correlated).

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Molluscs from Holocene Deposits of the Khasan Lake Region in Southern Primor'e

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Abstract – The southern part of the Primorskii region is one of the best studied areas of the Russian Far East. Abundant information related to various aspects of Holocene stratigraphy and paleogeography has been collected. Moreover, data on spore and pollen spectra, foraminifer and diatom complexes, and radiological dating (Korotkii *et al.*, 1980, 1988) are the predominant basis of the stratigraphic schemes and reconstructions of the environment. Holocene malacofauna has been studied in the greatest detail in the central part of the northwestern coast of the Sea of Japan (Evseev, 1981) and on the shelf of southern Primor'e in the Bay of Vostok (Evseev, 1975, 1976).

Borehole drilling in the Khasan District of the Primorskii region near the coastline of the Bay of Pos'et was carried out in March, 1991 by the Coastal Research Center of the Far East State University. Macrofauna (molluscan shells) with samples taken at 5 cm intervals from the drill cores of 4 boreholes have been passed on to me for analysis (37 samples). The results of this study are discussed below.

The Doritsine (Lotos) Lake. The borehole was drilled at the southern part of the lake (water depth: 1.4 m); the following beds were identified throughout the section (the thicknesses will be given in meters from now on):

1. Organogenic brown ooze	4.65
2. Silty sand	0.25
3. Bluish-black plastic mud	4.00
4. Bluish-black, sandy silt	0.55
5. Greenish-black, slightly silty sand	0.10
Total:	9.55

An oyster bed with abundant *Crassostrea gigas* (Thunberg) was reached in the interval of –8.9 - –9.1 m; two species of gastropods, *Rapana venosa* Valenciennes and *Tritia acutidentata* (Smith), and two species of bivalves, *Trapezium liratum* (Reeve) and *Macoma* sp., were found together with the oysters. The assemblage that was revealed apparently existed under shallow-water conditions, in a warmed up, and slightly desalinated bay.

The Sak-Pau (Zarechnoe) Lake. This section (water depth: 1 m) is composed of:

1. Semifluid, brown mud, grading into semifluid, plastic mud	3.55
2. Soft plastic clay, black with bluish tint	7.85
3. Greenish-grey sand	0.15
Total:	11.55

Molluscs are absent in the upper part of the section. However, *R. venosa Valenciennes*, the inhabitant of semi-closed bays with a calm hydrodynamic regime, was found in the interval of –9.8 - –9.85 m.

The Lebedinyi Bay. This desalinated bay (lagoon) is connected with Expedition Bay in the Gulf of Pos'et. The drilling penetrated the following beds (water depth: 1.25 m):

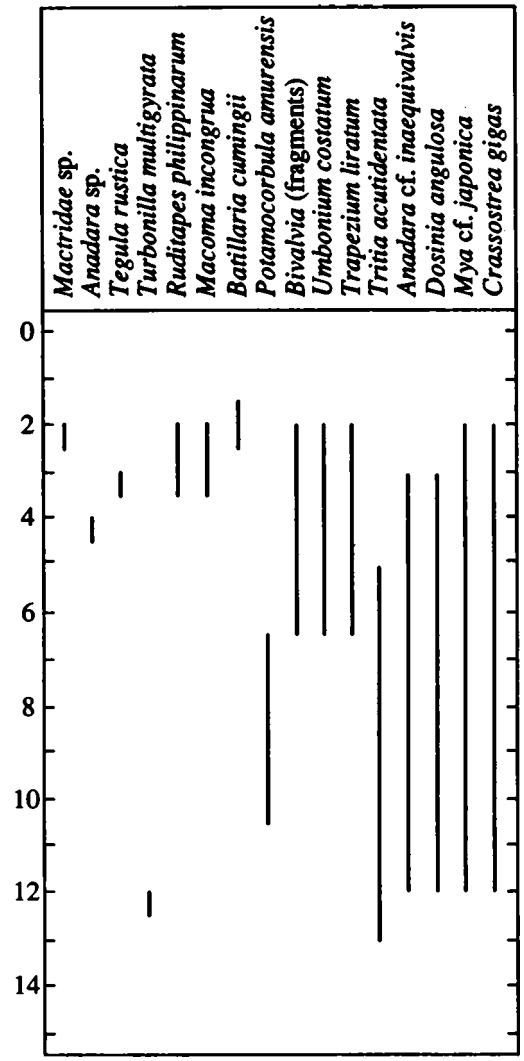
1. Silty, bluish-black sand	4.10
2. Slightly silty, light-gray sand	0.20
Total:	4.30

The bivalve bed of the *Corbicula* cf. *japonica* Prime, a typical representative of the southern Primor'e estuary complex, was found in the interval of –3.25 - –3.35 m. Well-preserved shells were found with traces of, or almost completely preserved periostracum. A single specimen of *Tritia acutidentata* (Smith) was also found in this interval.

The Lagoon of Tal'mi. This large, shallow-water (maximal depth: 1.5 - 2 m) lagoon is joined with the Sivuch'ya Inlet through a narrow watercourse. A borehole was drilled in the northern part of the lagoon and the following beds were identified (water depth: 1 m):

1. Dark-gray, fine-grained, silty sand with shell detritus	2.50
2. Dark-gray silt, alternating with thin interbeds of sand and clay	0.90
3. Black plastic, silty clay with thin interbeds of gray silt	10.05
4. Yellowish-gray, slightly silty, fine-grained sand	1.55
Total:	15.00

Macrofauna from the 1 m thick, uppermost layer was not sampled. In general, molluscs are represented in all these beds by 15 species, including 5 gastropods and 10 bivalves. The distribution of species (the figure)



Distribution of bivalve and gastropod molluscs in the borehole of the Lagoon of Tal'mi.

shows a definite succession in the colonization of the lagoon by molluscs. Judging by the absence of shells in the underlying sands (bed 4), the first molluscan colonies appeared where the modern lagoon is situated at the beginning of the sedimentation of the thick bed of silty clay. *Tritia acutidentata* (Smith) was found within the interval of –12.45 - –12.55 m, and *Turbonilla multigyrata* Dunker appears at the interval of –12.20 - –12.25 m. The latter are common marine gastropods, which live in southern Primor'e inside well-protected, warmed up bays. Two shell assemblages can be identified in the upper part of the section. *Crassostrea gigas* (Thunberg), found in 20 out of 33 samples containing macrofauna, is a common species of the first complex (–2 - –12 m). Characteristic forms of oyster reefs and banks, *Mya* cf. *japonica* Jay, *Tritia acutidentata* (Smith), and *Trapezium liratum* (Reeve), were found together with *C. gigas* (Thunberg). The second assemblage (above –2 m) includes *Macoma incongrua* (Martens), *Ruditapes philippinarum* (Adams et Reeve), and *Tegula rustica* (Gmelin), which lived here during the final phase of the oyster colony's existence. At that time, *Batillaria cumingii* (Crosse), which was recorded at an interval of –1.90 - –2.45 m, appeared in this region.

Thus, three stages of malacofauna formation were recognized. At the first stage, when marine water flooded the wide coastal plain, a shallow-water bay was formed in the area of the present-day Lagoon of Tal'mi. An intensive accumulation of silty sediments began forming, and molluscan colonies, with *Tritia acutidentata* (Smith) as the dominant species, appeared. At the second stage, an oyster bank with typical fauna developed on silt. This bank existed until dark-gray silt and sand (bed 2) began to accumulate. The last oysters disappear at approximately –2.4 m, whereas the boundary between beds 1 and 2 is at –2.5 m. A shallow-water, semiclosed bay existed in place of the lagoon during the period of the oyster bank's existence. This bay was slightly desalinated due to an inflow from the river, which is indicated by the presence of *Potamocorbula amurensis* (Schrenck), an euryhaline marine species in the sediments. These species prefer water with a reduced salinity. At the third stage, a change in the sedimentation regime took place, which is connected with a falling sea level and caused the shoaling of the bay. The break in the connection with the open sea resulted in the extinction of the oyster bed and the colonization of the lagoon by euryhaline marine and brackish-water molluscs. One of them, *Batillaria cumingii* (Crosse), still exists here; the others, such as *Macoma balthica* (L.), a mass form of present-day lagoon benthos, appeared at the latest satges of the basin's formation. This is revealed by a comparison of the list of recent malacofauna (Volova, 1974) with the specific composition of molluscs from the borehole.

The data obtained suggest the foliowing conclusions: The molluscan assemblage in the interval of –2 - –12 m is faunally uniform and may be correlated with the warm-water fauna found in sections of the coastal terraces and plains of southern Primor'e (Korotkii *et al.*, 1980; Evseev, 1981; Lutaenko, 1991) and on the Hokkaido Island coast (Ohshima *et al.*, 1972; Matsushima, 1984; and others). The biogeographical analysis of molluscs from this interval testifies to the prevalence of warm-water tropical-subtropical (23%), subtropical (23%), and subtropical-subboreal (46%) forms in the composition of the assemblage; a single, widely distributed boreal species (8%) is also present. Molluscs in the interval –2 - –12 m exhibit a significant similarity in specific composition to the rich fauna from the coast of the innermost part of the Bay of Ussuri, which I attribute to the Holocene climatic optimum (Lutaenko, 1988). The molluscan assemblage, which was studied earlier in the sediments of the terrace, and is about 4 m above the coast of the Bay of Tal'mi (Alekseev and Golubeva, 1980), has the same middle Holocene age. Hence, the thick clay bed, below the –2 - –3 m interval in the lagoon sediments, belongs to the Atlantic time, according to the Blytt–Sernander classification scheme; the overlying beds belong to the unsubdivided, late Holocene. The roof of the Atlantic

sediments can be ascertained by the disappearance of the thermophilic, tropical-subtropical *Anadara* cf. *inaequivalvis* (Bruguiere), now living in southern Primor'e and the thermophilic species *Dosinia angulosa* (Philippi) and *Trapezium liratum* (Reeve) in the interval of –3.25 - –3.30 m. Apparently, oyster beds continued to exist even during the Subboreal; however, the complete replacement of the assemblage in the late Holocene was connected with a progressively diminishing salinity due to a break in the connection with the open sea, rather than with climatic changes. Replacement of marine by lagoon facies can also be seen in the boreholes drilled in the Sak-Pau and Doritsine Lakes. Marine beds with shells occur here no higher than –8 - –10 m and accumulated during the period of the higher sea level, i.e., during the Atlantic time. According to A.M. Korotkii *et al.* (1980), the Atlantic sediments are a primary unit (in volume), filling ingressive lagoons along the river valleys and the bays of Primor'e. In general, they are most widely distributed in the coastal zone.

It should be noted that another but also thermophilic molluscan assemblage, characterized by the absence of tropical-subtropical species, (Bersenev *et al.*, 1983), was living in the open areas of the Bay of Pos'et during the Atlantic. This clearly indicates the distinct differentiation of sublittoral molluscan colonies in the southern Primor'e shelf during the middle Holocene. This difference was inherited by the modern malacofauna of this region.

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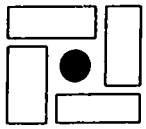

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